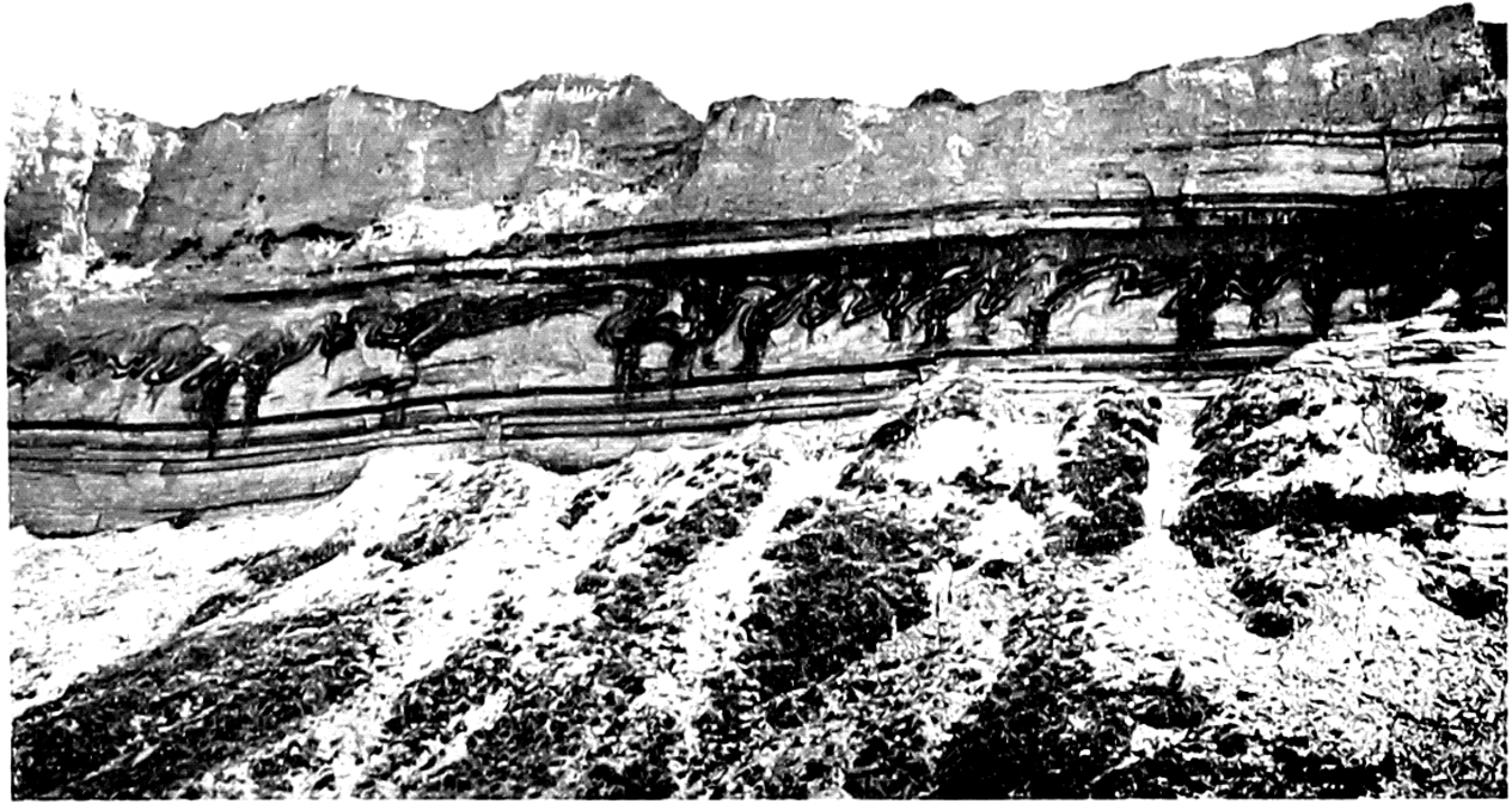
OUTLINES OF STRUCTURAL GEOLOGY

PLATE I



INTRAFORMATIONAL CONTORTED BEDS, AWA-AWA BEACH, NEW ZEALAND

The contorted beds consist of interbedded sandstone and siltstone, dipping at less than 1°. It is suggested by Ongley and Macpherson that these beds retained water, and moved under the pressure of the overlying beds.

(*Photo: M. Ongley.*)

OUTLINES OF STRUCTURAL GEOLOGY

by

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PREFACE

IN this book, I have aimed at presenting a brief, yet reasonably complete and well-documented summary of structural geology, with special reference to those aspects of the subject with which the field geologist should be acquainted. The nomenclature of certain sections of structural geology is at present in an unsatisfactory condition, and I have therefore given as full a synonymy as appeared to be necessary for the guidance of students in wider reading, which, it is hoped, the bibliographic references will facilitate.

It is regretted that, owing to difficulties connected with the geographical remoteness of Australia from the western countries, personal communication with all the authors whose figures have been reproduced, with or without modification, has not been possible. I am, however, especially indebted to Professor Hans Cloos, Mr. M. Ongley, and Mr. H. W. Fairbairn for their help in connexion with illustrations, and to Mr. E. J. Wayland for generously permitting me to make use of previously unpublished figures concerning the East African Rifts.

I am also happy to tender thanks to Prof. J. A. Bartrum, who kindly criticized parts of the preliminary MS.; to Dr. F. J. Turner, Dr. F. Coles Phillips and Mr. E. Ingerson, who advised me concerning petrofabric analysis; and to Mr. D. E. Thomas, of the Geological Survey of Victoria, who has discussed problems with me on many occasions. The final responsibility for all statements rests, however, with me.

My thanks are also due to Dr. O. M. B. Bulman of the Sedgwick Museum, Cambridge, for kindly reading the

final proofs on my behalf; and also to the following for permission to reproduce illustrations or use them as a basis for some of the diagrams in this work: The Editors of the *Geological Magazine* (Figs. 3, 10, 37), The Geological Society of America (Figs. 8, 43, 81), The Royal Society of Edinburgh (Fig. 9), The McGraw-Hill Publishing Co., Ltd. (Figs. 18, 19), Julius Springer Verlagsbuchhandlung, Berlin (Figs. 17, 22, 23, 41, 103, 104, 106), Edward Arnold & Co. (Figs. 29, 31, 35), Willibald Keller Verlagsbuchhandlung, Leipzig (Figs. 30, 79), The United States Geological Survey (Figs. 32, 80, 84), H.M. Stationery Office (Fig. 36), E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart (Figs. 52, 94), Cambridge University Press (Figs. 47, 85), John Wiley & Sons, New York (Figs. 51, 61), The American Association of Petroleum Geologists (Fig. 57), *The Journal of Geology*, Chicago (Figs. 71, 89A), Henry Holt & Co., New York (Fig. 83), Dietrich Reimer Verlag, Berlin (Fig. 92), Ernst Cloos and The National Academy of Sciences, Washington (Fig. 96).

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PREFACE TO THIRD EDITION

THE tenets of many branches of structural geology have been greatly modified by researches and discussions during the past decade, so much so that in some branches, as for instance, the mechanism of orogeny, a state of philosophic flux has developed that is perhaps even more remarkable than the *tectonique d'écoulement* which is the subject of much contemporary European work. If only because of the manifest insufficiency of hypotheses formerly regarded as conservative, some account of such and similar matters has been presented despite their admittedly controversial nature, especially since it is hoped thereby to avoid the inculcation of rigid notions that tend to restrict thought rather than to guide it along profitable lines of inquiry.

The Publishers having collaborated generously in providing for revision, opportunity has been taken to introduce some treatment of tectonic concepts, to the extent that a knowledge of these is thought to be essential to an understanding of structural features. At the other end of the scale, the chapter on Petrofabrics is retained despite its brevity, its chief value being regarded as the indication it gives that structural geologists may, with advantage, use a microscope.

It is a pleasure to acknowledge the assistance of colleagues and friends who have commented on this book or who have discussed problems with me, but I must take full responsibility for statements concerning the views of these and other geologists, to whose work specific reference is made in the text.

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Chapter I

NON-DIASTROPHIC STRUCTURES

1. *PRIMARY STRUCTURES OF SEDIMENTARY ROCKS*[1]

Stratification.—The arrangement of sedimentary rocks in layers is one of their most characteristic features. The various layers, which are separated by the *bedding* or *stratification planes*, are usually distinguishable one from another by differences in composition, texture, hardness, cohesion, or colour. A *bed* or *stratum* is a layer which is composed throughout of similar material, but thin layers showing minor variations in texture or other properties may be present in an individual bed. These minor layers are referred to as *laminae* if they are less than about half an inch in thickness, or *stratification layers* if they are thicker. Rocks can generally be split easily along the stratification planes, and in fissile shales the orientation of lamellar or rod-like minerals parallel to these planes produces a definite cleavage. The orientation of these minerals is believed to be caused by their rotation during the gravitational compaction of original clayey sediments, and by plastic flow parallel to the bedding,[2] so that the cleavage is not a primary structure and should be distinguished from the fine lamination of paper shales.

Facing of Strata.—It is convenient in dealing with highly

[1] For further general information on primary structures consult Twenhofel, W. H., *Treatise on Sedimentation*: London, 2nd edn., 1932, pp. 603–756. Andrée, K., 'Wesen, Ursachen und Arten der Schichtung': *Geol. Rundsch.*, Vol. 6, 1916, pp. 351–97. Tyrrell, G. W., *Principles of Petrology*: London, 3rd edn., 1934, pp. 196–202.

[2] Lewis, J. V., 'Fissility in Shale and its relations to Petroleum': *Bull. Geol. Soc. Amer.*, Vol. 35, 1924, pp. 557–90.

disturbed rocks, especially where overfolding is common, to refer to the determination of the original top of a bed, which is of fundamental importance in mapping, as the determination of its *facing*. Many criteria for recognizing top and bottom of sedimentary rocks, lava flows, and pyroclastic rocks have been used, and doubtless others will be forthcoming.[1] The subject is referred to in several different contexts in this book.

Grading of Sediments.—Frequently, there is a gradual variation, in a vertical direction, in the size of the grains of which a bed, lamina, or stratification layer is composed. The gradation within an individual layer is usually from coarser material below to finer above, a sharp break occurring between the fine material of the upper part of the lower layer and the coarse bottom material of the layer above. *Graded bedding*, as this textural variation is termed, may result from variation in the size of particles supplied to a sedimentary deposit, as in glacial varves, but more commonly in marine deposits it arises from the fractionation of mixed coarse and fine particles originally admixed in a water current. Fractionation may occur during the settling through still bottom waters of mixed sediments carried in a superficial current,[2] and Bailey has suggested that such detritus-laden currents may be caused by seaquakes, which give rise to so-called 'tidal waves' or *tsunamis*.[3] Again, Kuenen and Migliorini[4] have demonstrated that fractionation occurs in turbidity currents, and it is very likely that many marine successions in which argillaceous sandstones (greywackes) commonly grade to shale or mudstone, have been deposited by such currents.

Water carrying suspended sediment is able to flow down

[1] See especially Shrock, R. R., *Sequence in Layered Rocks*: New York, 1948.

[2] The gradation exhibited by coarse current bedded strata (see Wills, L. J., *The Physiographical Evolution of Britain*: London, 1929, p. 119) is a distinct type, and is not generally classed as graded bedding.

[3] Bailey, E. B., 'New Light on Sedimentation and Tectonics': *Geol. Mag.*, Vol. 66, 1930, pp. 77–92; 'Sedimentation in Relation to Tectonics': *Bull. Geol. Soc. Amer.*, Vol. 47, 1936, pp. 1713–26.

[4] Kuenen, Ph. H., and C. I. Migliorini, 'Turbidity Currents as a Cause of Graded Bedding': *Journ. Geol.*, Vol. 58, 1950, pp. 91–127.

gentle bottom-slopes in standing water, because of the relatively high specific gravity of the water-sediment mixture, which constitutes a turbidity current. High-density currents can transport fragments up to pebble size, presumably for long distances, and experiments show that the ultimate deposits are graded. Turbidity currents are known to be formed in lakes and reservoirs by flooded streams, but it is suggested that they may also be produced on a large scale in the sea, chiefly by the

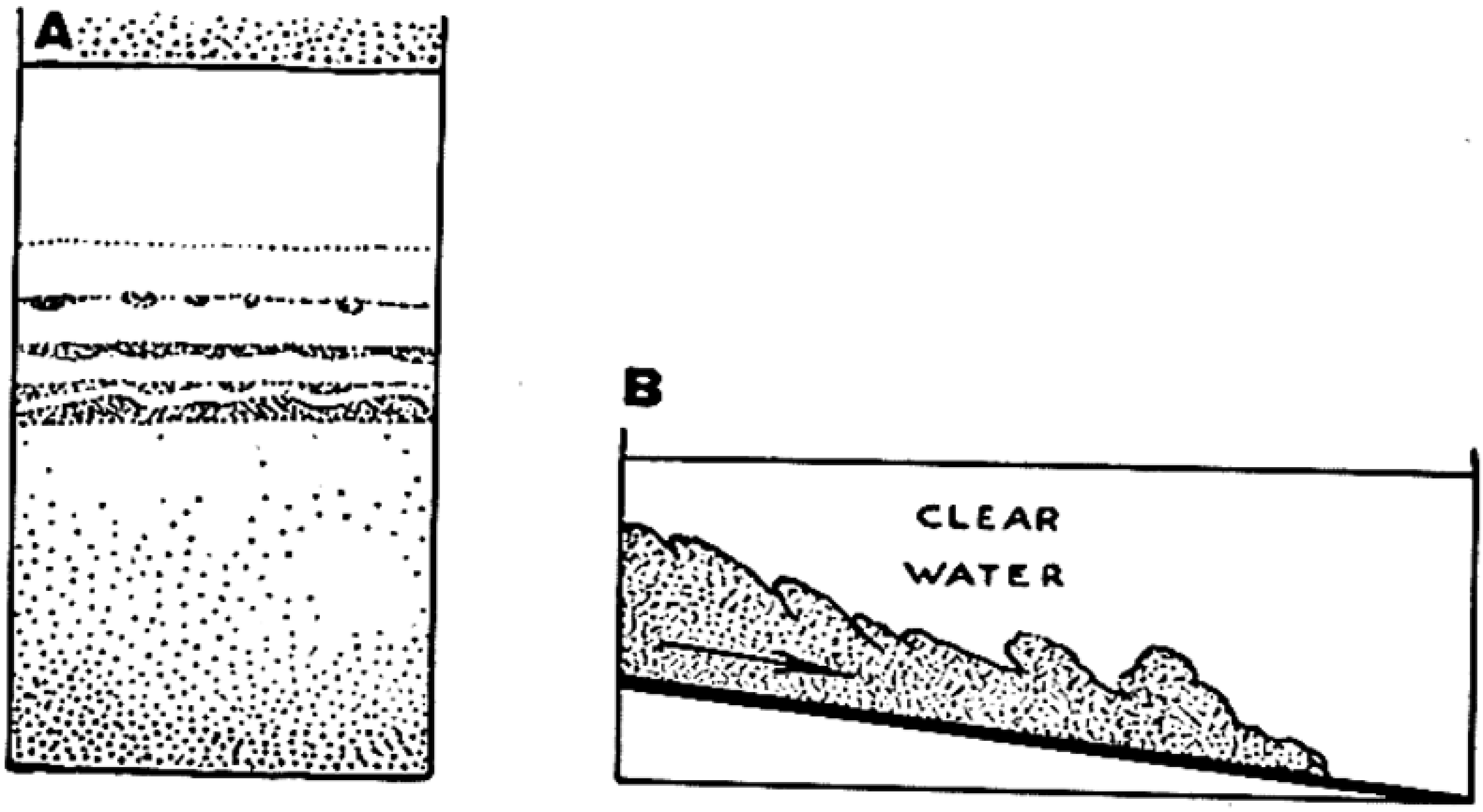


FIG. 1.—GRADED BEDDING

A. Compound graded bed with arenaceous base, laminated zone (with current ripple mark, etc.) and argillaceous top.
B. Turbidity current in water tank, showing secondary waves.

slumping of previously deposited sediments. Once formed, a turbidity current may erode soft sediments, thus adding to its importance as a transporting agent.

Although it has been stated that current bedding and graded bedding are formed in different environments[1] the regular transition from argillaceous sandstone to shale or mudstone in a thick graded bed may be modified by a laminated zone between the arenaceous base and the argillaceous top, in which the sandy layers are commonly both current-bedded and ripple marked. Such features in the thick Ordovician succession in

[1] Bailey, E. B., *op. cit.*, 1930, 1936.

Victoria[1] are thought to have originated from subsidiary undulations or waves in turbidity currents (Fig. 1).

If used with discretion, graded bedding can be very useful in mapping,[2] but it is important to note that, owing to the fact that the gradation in coarse sediments may sometimes be the reverse of that above described,[3] valid evidence as to the order of superposition can only be obtained from a succession of fine-grained graded beds, giving concordant results.

Initial Dip.—The original attitude of stratification planes is usually approximately horizontal, but cross-bedded deposits (see pp. 5–7) and sediments laid down on sloping surfaces have an *initial dip*, which is imparted to them during deposition and not by later disturbances. The maximum angle of initial dip depends upon the angle of rest of the sediments under the conditions obtaining during their deposition, and may be as much as 43° in the case of sand laid down in quiet water, though finer grades of sediment have lower angles of rest.[4]

Sediments with moderate to high initial dips are to be expected at steep sites of deposition developed from any cause, and are common around coral reefs and volcanoes, also against fault scarps and buried hills (see pp. 67–8, 85–6). In field work the possibility should be remembered that an observed dip may be initial and not due to later tilting or folding. Initial dips are modified by compaction of the sediments and by earth movements,[5] and this should be borne in mind in making deductions concerning the original conditions of sedimentation.

Discordant Bedding.—The layers in a stratified deposit laid

[1] Hills, E. S., and D. E. Thomas, 'Fissuring in Sandstones': *Journ. Geol.*, Vol. 40, 1945, pp. 51–61.

[2] See Read, H. H., 'Dalradian Rocks of the Banffshire Coast': *Geol. Mag.*, Vol. 73, 1936, pp. 468–76, for an account of the metamorphism of graded beds.

[3] Cooke, H. C., 'Anomalous Grain Relationship in the Caldwell Quartzites of Thetford District, Quebec': *Proc. and Trans. Roy. Soc. Canada*, Sect. 4, Vol. 25, 1931, pp. 71–4.

[4] Draper, M. B., quoted in Twenhofel, W. H., *Treatise on Sedimentation*: 2nd edn., 1932, pp. 604, 605.

[5] Wilson, I. F., 'Buried Topography, Initial Structures and Sedimentation in Santa Rosalia Area, Baja, California, Mexico': *Bull. Amer. Assoc. Petrol. Geol.*, Vol. 32, 1948, Pt. 2, pp. 1762–1807.

down under conditions of quiet sedimentation are sensibly parallel with each other, and, with the exceptions noted above, are approximately horizontal. In detritus that is deposited rapidly from heavily laden wind or water currents, on the other hand, the stratification planes are often inclined at the angle of rest of the sediments, or regularly waved showing a sigmoidal curve in cross-section, or irregularly waved and inclined. The minor stratification planes in such deposits are oblique to the major bedding planes that separate the larger units in the sedimentary series, and the bedding is said to be *discordant.*[1] The terms *current bedding*, *cross bedding*, *false bedding*, *oblique bedding* and *inclined bedding* are also used more or less synonymously to designate this structure, but current bedding is perhaps best used in connexion with deposits laid down by water, and exhibiting waved stratification planes. Other types of discordant bedding may be named according to their mode of origin, if determinable, as varieties of cross bedding, e.g. aeolian cross bedding, deltaic cross bedding, etc.

Discordant bedding occurs in sand dunes, bars, beaches, deltas, fluviatile deposits, lacustrine and marine deposits subjected to moderate or strong current action, and is also exhibited on a small scale in ripple-marked strata.[2]

In deltas there are three distinct sets of deposits, the *top-set*, *fore-set*, and *bottom-set beds* (see Fig. 2). The top-sets have a low initial dip, due to their being deposited on the sloping surface of the subaerial part of the delta. The fore-sets, which are composed of material dropped over the outer edge of the delta, have an initial dip equal to the angle of rest of this material. The bottom-sets or *pro-delta clays* represent the finer detritus spread out over the floor of the sea or lake in which the delta

[1] Tyrrell, G. W., *The Principles of Petrology*: London, 3rd edn., 1934, pp. 198, 199.

[2] For general accounts of discordant bedding, see Andersen, S. A., 'The Eskers and Terraces in the Basin of the River Susa': *Geol. Surv. Denmark*, II Raekke, 1931 (in English). Grabau, A. W., *Principles of Stratigraphy*: New York, 1st edn., 1913, pp. 701–6. Thompson, W. O., 'Original Structures of Beaches, Bars, and Dunes': *Bull. Geol. Soc. Amer.*, Vol. 48, 1937, pp. 723–52. Twenhofel, W. H., *Treatise on Sedimentation*: London, 2nd edn., 1932, pp. 618–23.

was formed; they have a low initial dip, and merge with the base of the fore-sets in a gentle curve. The upper ends of the fore-sets are either truncated by the top-sets or join with them in an abrupt curve. In the latter case the bedding planes are continuous throughout the top-, fore-, and bottom-set beds, and exhibit a double curve in longitudinal cross-section.

The deposition of coarse sediments by torrential streams takes place by a 'dumping' process analogous to the building

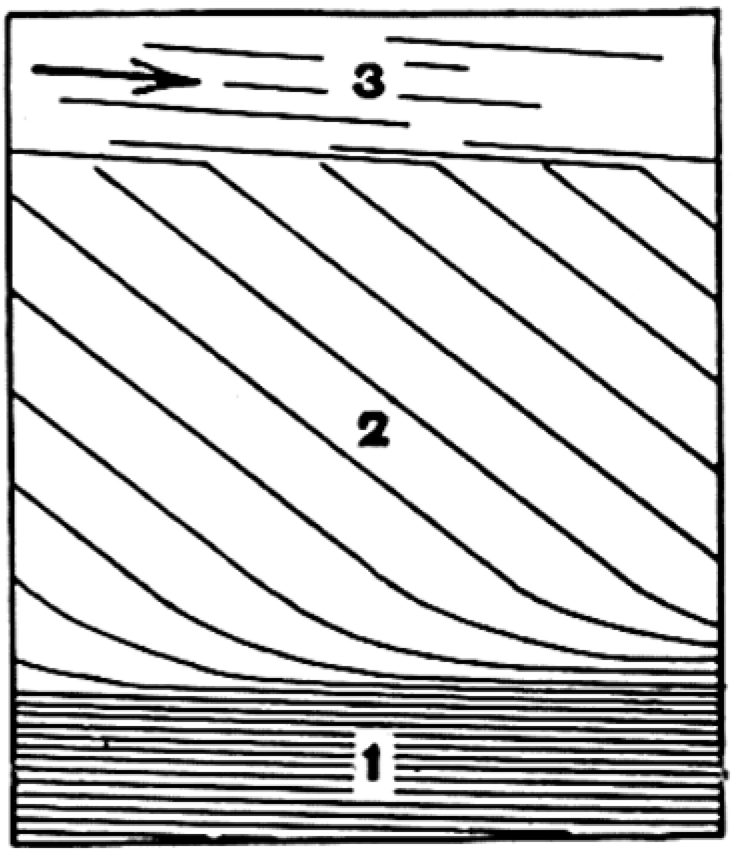
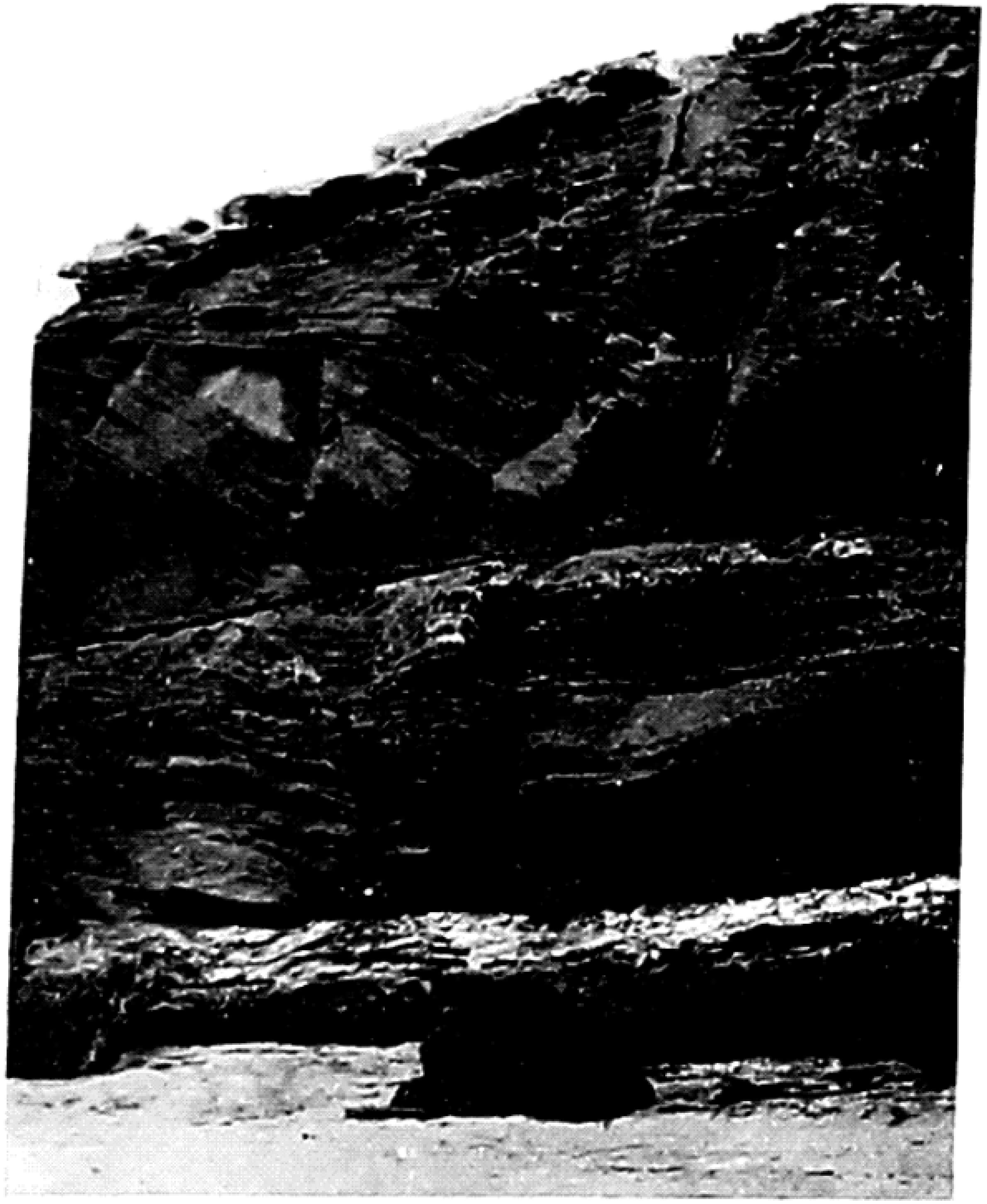


FIG. 2.—LONGITUDINAL CROSS-SECTION OF DELTAIC DEPOSITS

The arrow indicates the direction of supply of detritus. 1, Bottom-set beds; 2, fore-set beds; 3, top-set beds.

of delta forests, but the stratification is usually very irregular, and the top-set and bottom-set beds sub-ordinate or absent (*torrential cross bedding*).

In the finer grades of sediment deposited from heavily laden currents, the stratification planes are, with appropriate relationships between current strength and rate of supply of detritus to any point, sigmoidally curved (see Fig. 3). The upper portions of these curved stratification planes may, however, be eroded off if the current strength increases, or if the supply of sediment decreases, so that their upper ends are truncated by the next layer to be deposited. Since this truncation can only occur at the top of a stratum, we have here, as with deltaic cross bedding, a means of determining the correct order of



A. Aeolian cross bedding in Pleistocene dune sandstone, Barwon Heads, Victoria. The fore-sets in the highest dune are locally truncated at the base and gently curved at the summit.

(Photo: E. S. H.)

B. Flow wrinkles on the underside of a basalt flow covering a steep scoria slope, Mt. Porndon, Victoria. The flow is inclined away from the observer. Lava stalactites are shown hanging from the roof of a small lava tunnel at the base of the flow. *(Photo: A. F. McQueen)*

superposition of a sedimentary series exhibiting discordant bedding.[1] The rule is to look for the truncation of the inclined laminae, which is always at the top of the bed, rather than for the concave surfaces of the stratification planes, which may face either up or down (see Fig. 3). Descriptions of the use of this method in the field may be found in the papers cited below.[2]

In *aeolian* or *dune bedding*, the fore-sets may be locally continuous with the 'top-sets', and show no curvature at the base. This arrangement is found in dunes which are built up on both the windward and lee slopes (Pl. II, A), and it bears a certain

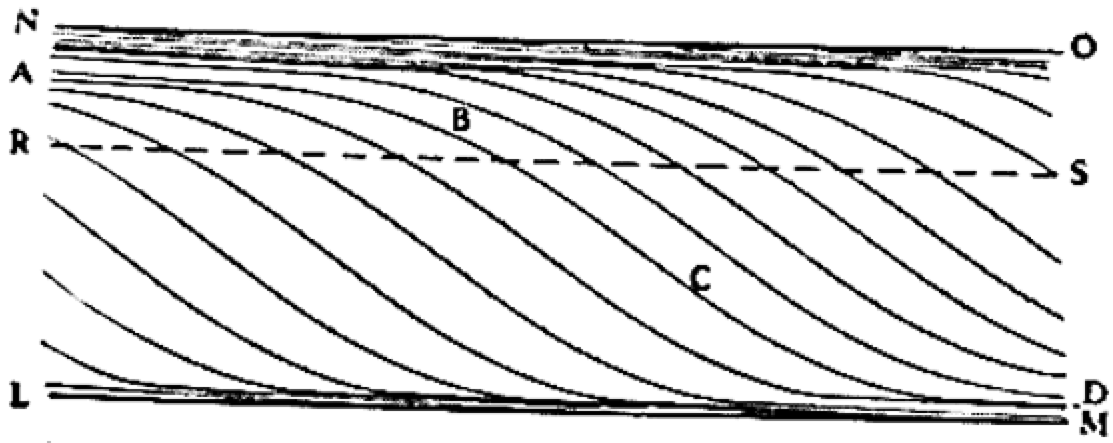


FIG. 3.—A COMPOSITE STRATUM WITH TOP *NO* AND BASE *LM*, COMPOSED OF LAYERS WHICH POSSESS THE COMPLETE CURRENT-BEDDING CURVE, *ABCD*

(After Bailey, 1930)

Penecontemporaneous erosion, by removing the part above *RS*, causes truncation of the remaining stratification layers.

resemblance to inverted deltaic cross bedding. Failure to recognize its origin might lead to error in the application of the above rule.

Ripple Mark.—Ripple mark[3] on sediments is caused by the

[1] Bailey, E. B., 'New Light on Sedimentation and Tectonics': *Geol. Mag.*, Vol. 67, 1930, pp. 77–92.

[2] Vogt, T., 'On the Chronological Order of Deposition of the Highland Schists': *Geol. Mag.*, Vol. 67, 1930, pp. 68–73. Bailey, E. B., 'West Highland Tectonics: Loch Leven to Glen Roy': *Quart. Journ. Geol. Soc.*, Vol. 90, 1934, pp. 462–525. Allison, A., 'The Dalradian Succession in Islay and Jura': *Quart. Journ. Geol. Soc.*, Vol. 89, 1933, pp. 125–44. Merritt, P. L., 'Seine-Coutchiching Problem': *Bull. Geol. Soc. Amer.*, Vol. 45, 1934, pp. 333–74.

[3] For a good summary of the literature relating to recent and fossil ripple mark, see Twenhofel, W. H., *Treatise on Sedimentation*: 2nd edn., 1932, pp. 632–68. Also Bucher, W. H., 'On Ripples and Related Sedimentary Surface Forms and their Palaeogeographic Interpretation: *Amer. Journ. Sci.*, Vol. 47, 1919, pp. 149–210, 241–69.

movement of air or water over the unconsolidated surface layers of incoherent deposits. The movements may be either continuous in the one direction (currents), or oscillatory, the latter being produced by waves in standing bodies of water shallower than wave base. A simple classification of ripple mark into *current ripple mark* and *wave* (or *oscillation*) *ripple mark* depends upon this distinction. In the description of ripple mark, *amplitude* refers to the elevation of a ripple crest above the neighbouring troughs, *wave length* to the distance from crest to crest, and the *ripple index* is the wave length divided by the amplitude (see Fig. 4).

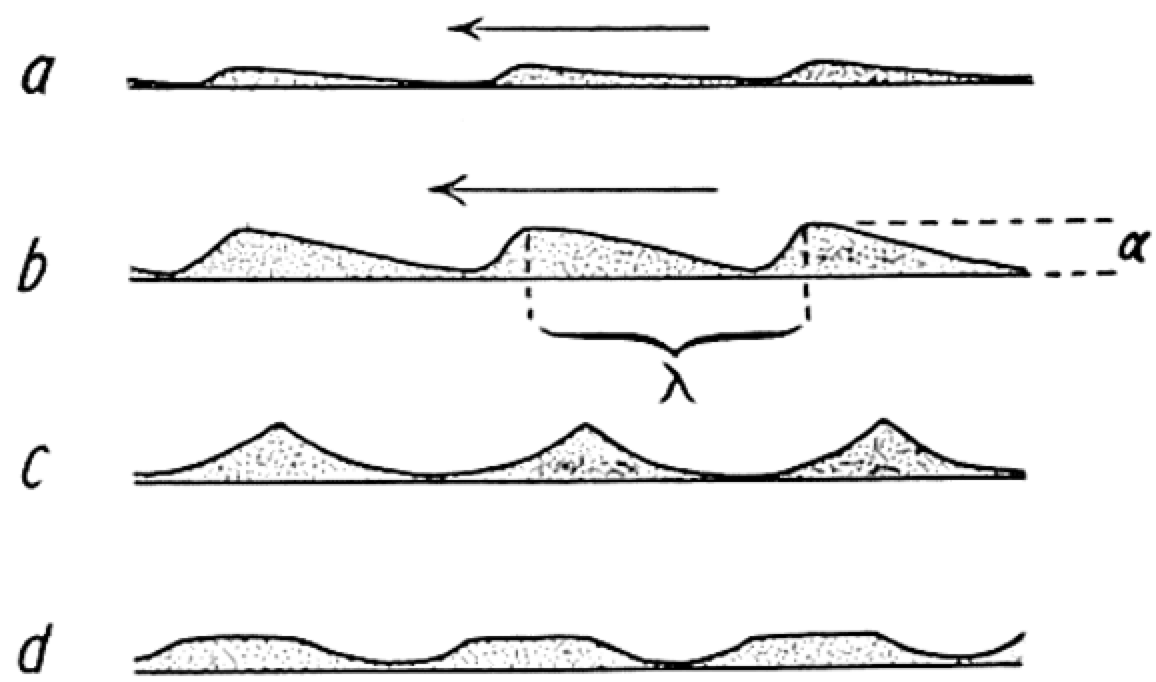


FIG. 4.—TYPES OF RIPPLE MARK ON SEDIMENTS

a, aeolian current ripple mark; *b*, sub-aqueous current ripple mark. Note the greater amplitude, as compared with the wave length, of the sub-aqueous ripples. *c*, wave or oscillation ripple mark; *d*, wave ripple mark with eroded crests. λ, wave length, α, amplitude. The arrows indicate the direction of the current in *a* and *b*.

Current ripples are asymmetrical and move in the direction of the current like miniature sand dunes, the gentler slopes facing upstream and the steeper slopes downstream. The crests of well-formed current ripples are sharp, and the troughs rounded. With increasing stream velocity current ripples are destroyed, and the surface of the sediments is thrown into large wave-like ripples, more or less symmetrical, which are not commonly preserved in the geological column. Aeolian current ripples, developed on wind-blown sand, resemble sub-aqueous current ripples but are said to differ from them in having

a relatively greater wave length as compared with their amplitude[1] (see Fig. 4). It is not yet certain, however, that this difference always obtains.

In water that is shallow enough for the deepest parts to be affected by movements due to waves, the water moves in an oscillatory manner over the bottom as each wave passes, heaping up the sand into stationary parallel ridges with sharp crests separated by rounded troughs. Since the sharp crests of well-formed wave ripples are easily distinguishable from the smoothly rounded bottoms of the troughs, the upper surface of a bed showing this type of ripple mark may readily be determined. Some examples, however, have somewhat rounded crests, and it is then impossible to use the ripple mark as a stratigraphical criterion. Also, where ripple-marked sands occur within the tidal range, the retreating waves, aided by the wind, erode off the sharp crests, and in overturned beds these eroded crests might be confused with shallow troughs (see Fig. 4). The safest course in field work is to search for normal examples, one set of wave ripples with sharp crests sufficing as valid evidence in determining the top of a bed. Perfectly formed current ripples can also be used in the same way.

Current ripple-marked sediments generally exhibit small-scale current bedding, and where the wave length is constant in successive laminae but crests and troughs gradually shift position in a downstream sense, an apparent crossbedding dipping upstream is produced. This appearance is intensified if there is preferential deposition of particles of different size and shape in the troughs and in the steep downstream slopes of the ripples (Fig. 5).[2]

Compaction, and differential plastic deformation modify the form and internal structure of entombed ripple mark. Although such effects may make the reading of facing more

[1] Kindle, E. M., 'Recent and Fossil Ripple-mark': *Geol. Surv. Canada, Mus. Bull.* 25, 1917, 56 pp.

[2] See McKee, E. D., 'Some Types of Bedding in the Colorado River Delta': *Journ. Geol.*, Vol. 47, 1939, pp. 64–81. Twenhofel, W. H., *op. cit.*, 1932. A similar feature in oscillation ripple mark is mentioned by Gilbert, G. K., 'Ripple marks and Cross Bedding': *Bull. Geol. Soc. Amer.*, Vol. 10, 1899, pp. 135–40.

difficult, they are of interest since they afford some guide to the kind of internal movement that has affected strata, since the original form of the ripples and their associated current bedding may generally be approximately gauged.[1]

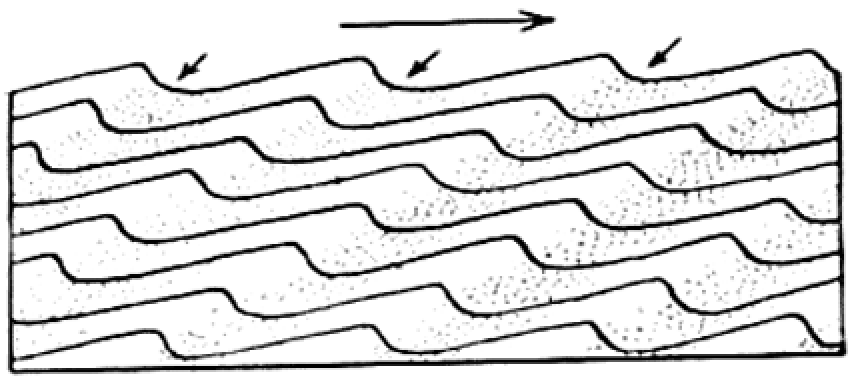


FIG. 5.—APPARENT CROSS BEDDING, WITH PSEUDO-STRATIFICATION INCLINED IN THE DIRECTION OF THE SMALL ARROWS, PRODUCED BY THE SUPERIMPOSITION OF CURRENT RIPPLE-MARKED LAYERS

The current flowed in the direction of the long arrow. Coarse particles, which are concentrated in the troughs of the ripples, are shown by stippling.

Rain, Drip, and Hail Impressions.—The impressions formed on soft muds by raindrops, drips from trees, and hailstones, may be preserved if conditions are favourable. Drip impressions and those of vertically falling raindrops are circular and bounded by ridges of equal height all round. Raindrops, however, do not usually fall vertically, and they therefore generally leave ovoidal impressions, each having a raised edge, highest on the side opposite to that from which the rain came. Hail impressions are ovoidal, and deeper and less regular than rain imprints. The upper surfaces of all these imprints being concave, and the casts on the lower surface of the overlying bed convex, the top of a bed showing fossilized imprints can be recognized.

Mud Cracks (Sun Cracks, Shrinkage Cracks).—With rare exceptions, mud cracks develop only when coherent muddy sediments are exposed to a dry atmosphere for a considerable time. If water carrying suspended matter again rapidly covers the cracked mud, or if wind-borne material is deposited over it, the cracks become filled with sediment of a different character, thus preserving them in a readily recognizable form. The

[1] See McKee, E. D., *op. cit.*, 1939

downward tapering of the cracks gives an indication of the base of the sedimentary series[1] (Fig. 6).

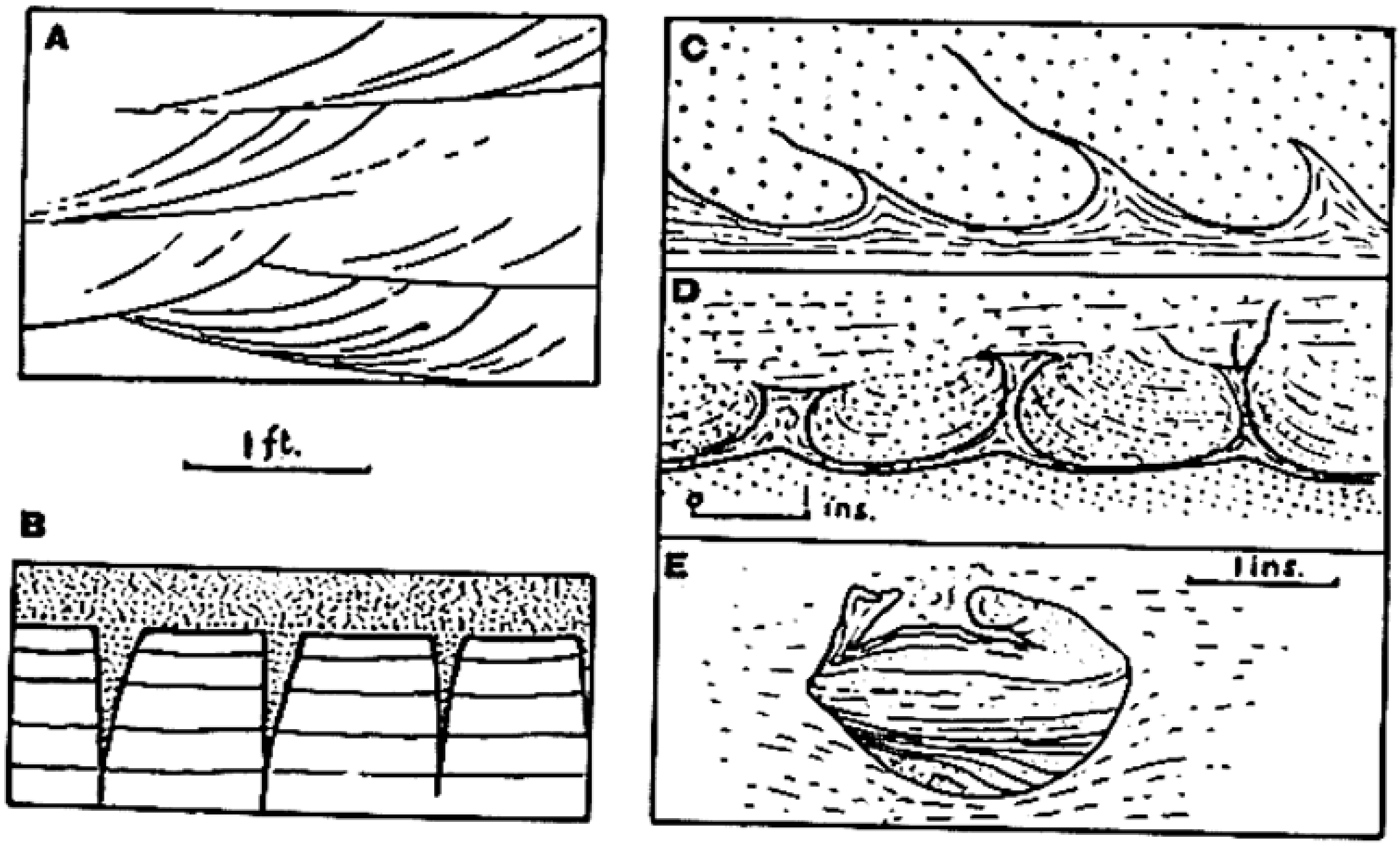


FIG. 6.—CRITERIA FOR FACING OF SEDIMENTS

A, Current bedding; B, mud cracks; C, Basal deformation of sandstone, with indications of lateral sliding; D, basal deformation of sandstone showing shale injections over ripple crests; E, 'Pseudo-nodule' of sand in shale. The top is upward in each example.

Included Fossils.—The attitude of included fossils can sometimes be used to indicate the upper stratigraphical surface of a bed. Not only can a knowledge of the mode of life of the organisms be applied to the problem, as for example with colonial corals, whose upper and lower surfaces are readily recognizable and which are often preserved in their original growth position,[2] but redistributed shells can also yield information.

[1] For a summary of the literature concerning shrinkage cracks and the various kinds of imprints, tracks, and trails, of which only those likely to prove useful in structural field work are treated here, see Twenhofel, W. H., *Treatise on Sedimentation*: London, 2nd edn., 1932, pp. 669–92; also 'Impressions made by Bubbles, Rain-drops, and other Agencies': *Bull. Geol. Soc. Amer.*, Vol. 32, 1921, pp. 359–72. Kindle, E. M., 'Some Factors influencing the Development of Mud Cracks': *Journ. Geol.*, Vol. 25, 1917, pp. 135–44. Abel, O., *Vorzeitliche Lebensspuren*: Jena, 1935.

[2] For a useful general account of the invertebrates see Twenhofel, W. H., and R. R. Shrock, *Invertebrate Palaeontology*: 2nd impn., New York, 1935.

Concavo-convex shells, for example, tend to be turned over by currents and waves so that their convex surfaces face upwards, and in shell banks subject to wave or strong current action the great majority of shells assume their attitude. After settling through quiet waters, on the other hand, shells come to rest with their concave surfaces uppermost, and it is necessary to bear this distinction in mind in the interpretation of fossil shell beds.

It has also been noted that internal moulds formed by the entry into the shells of fine-grained mud from the surrounding sediments, are not always fully developed, but may only partially fill the shell cavities. In such instances, the flat upper surfaces of the partial moulds are parallel with the original bedding planes, and face the stratigraphical upper surface.[1]

Penecontemporaneous Structures.—Where a sandstone rests on an argillaceous bed, the bottom of the sandstone and the underlying bed may be mutually involved in small-scale structures, in which the stratification in both beds is deformed (Fig. 6). Generally the base of the sandstone exhibits downwardly convex lobes, with sharp-crested projections of argillaceous material between them, which may extend as laminae for several inches into the sandstone. Such structures are due to the upward squeezing of the highly plastic argillite into the overlying sand while both were unconsolidated, and they often reveal, by a general asymmetry, slight differential slip of a few inches between the beds (Fig. 6C). Other injections are symmetrical (Fig. 6D) and resemble miniature structures produced experimentally in imitation of salt domes.[2] Hadding[3] has attributed asymmetrical lobation of sandstone bases to sub-

[1] Cullison, J. S., 'Origin of Composite and Incomplete Internal Moulds and their Possible Use as Criteria of Structure': *Bull. Geol. Soc. Amer.*, Vol. 49, 1938, pp. 981–8.

[2] Nettleton, L. L., 'Fluid Mechanics of Salt Domes': *Bull. Amer. Assoc. Petrol. Geologists*, Vol. 18, 1934, p. 1175. Dobrin, M. B., 'Recreating Geological History with Models': *Journ. Applied Physics*, Vol. 10, pp. 360–71.

[3] Hadding, A., 'On Subaqueous Slides': *Geol. Fören. Stockh. Förh.*, Vol. 53, 1931, pp. 377–93.

marine slumping, and Sorby[1] and Lamont[2] have described somewhat similar features as a kind of ripple mark called *anti-dune*, but where successive stratification planes are involved in the structures they may confidently be interpreted as post-depositional. They have been called *flow cast* or *basal sandstone deformations*.[3] Certain examples of balled-up sandstones or *pseudonodules* are of somewhat similar origin. These are isolated sand-rolls enclosed in a series along a particular horizon in shale or mudstone: each ball shows a rounded base with upturned or inrolled edges. Deformed stratification in the sand and in the enclosing shale shows that the sand sank into the shale, while the latter welled up between isolated sand-masses. Originally the sand may have been a lamina, probably ripple-marked, or a series of incomplete ripples as small lenses. Examples in the Ordovician rocks of Victoria (Fig. 6E) and artificial examples in the slime-dumps of mines support the interpretation of Macar and Antun[4] that balling-up took place soon after deposition.

On the other hand, sand balls may also occur haphazardly jumbled in mudstone, in beds that have clearly undergone considerable displacement in the soft-rock stage, either by slumping or by squeezing under load.[5]

Flow-casts and undisturbed pseudo-nodules may be used to indicate facing, since the rounded surfaces in both are directed downwards.

[1] Sorby, H. C., 'On the Application of Quantitative Methods to the Study of the Structure and History of Rocks': *Quart. Journ. Geol. Soc.*, Vol. 64, 1908, pp. 171–233.

[2] Lamont, A., 'Contemporaneous Slumping and other Problems at Bray Series, Ordovician, and Lower Carboniferous Horizons, in County Dublin': *Proc. Roy. Irish Acad.*, Vol. 45, Sect. B, No. 1, 1938.

[3] Shrock, R. R., *Sequence in Layered Rocks*: New York, 1948, p. 156. Hills, E. S., 'The Silurian Rocks of the Studley Park District': *Proc. Roy. Soc. Vict.*, Vol. 53, 1941, pp. 167–91. Hills, E. S., and D. E. Thomas, 'Fissuring in Sandstones': *Econ. Geol.*, Vol. 40, 1945, pp. 51–61.

[4] Macar, P., and P. Antun, 'Pseudo-nodules et glissement sous-aquatique dans l'Emsien inferieur de l'Œsling': *Bull. Soc. Géol. Belg.*, No. 4, Vol. 73, 1950, pp. 121–49. We are indebted to Dr. D. E. Thomas for reference to the Victorian occurrences mentioned.

[5] Kuenen, P. H., 'Slumping in the Carboniferous Rocks of Pembrokeshire': *Quart. Journ. Geol. Soc.*, Vol. 104, 1949, pp. 365–85.

Indications of the correct order of superposition of beds can also be obtained at local unconformities, where pebbles of older rocks are included in younger, and from scour channels, which are restricted to the lower beds and filled in with younger deposits. The curving of strata beneath pebbles has also been used in the solution of involved structures.[1] The use that may be made of secondary structures such as cleavage and drag folding will be indicated in later chapters.

2. *NON-DIASTROPHIC DEFORMATION*

Rock deformation is not always due to the direct intervention of diastrophic forces.[2] Differential compaction, the drag of moving ice, subaqueous slumping, and collapse under gravity can all cause structures which may simulate those resulting from diastrophism, and neglect of their possible effects may lead the field geologist astray.

Differential Compaction.—The expulsion of water from sediments, and the close packing of the particles by the load of superincumbent rock, result in a considerable shrinkage in volume. In clays this shrinkage may be as much as 75 per cent or more of the original volume of the deposit,[3] though in sandstones and limestones it is much less, Sorby giving the figure as less than 25 per cent. If lenticular beds of less compactable material, such as sand or reef limestone, are present in a series of clays, the original attitude of the stratification planes in the clays is altered by compaction. Originally horizontal clay beds assume a tilt away from the interbedded sand lenses or reefs, owing to the fact that the shrinkage in a given thickness of clays

[1] Collins, W. H., 'The Geology of the Gowganda Mining Division': *Mem. Geol. Surv. Canada*, No. 33, 1913, Pl. 3, Fig. 1, p. 84.

[2] The term diastrophism denotes the processes connected with major deformations of the earth's crust, such as mountain-building, or the elevation, depression, or warping of crustal blocks.

[3] Sorby, H. C., 'On the Application of Quantitative Methods to the Study of the Structure and History of Rocks': *Quart. Journ. Geol. Soc.*, Vol. 64, 1908, pp. 171–233. Hedberg, H. D., 'Gravitational Compaction of Clays and Shales': *Amer. Journ. Sci.*, Ser. 5, Vol. 31, 1936, pp. 241–87 (with good bibliography). Jones, O. T., 'The Compaction of Muddy Sediments': *Quart. Journ. Geol. Soc.*, Vol. 100, 1944, pp. 137–60.

between the lenses is greater than that of an equal thickness of clay with interbedded sand or limestone (see Fig. 7).

Similar effects result if the clays rest upon an uneven surface, the thick deposits in the hollows compacting more than the thinner deposits over the elevations, so that structural domes form over the elevations, and basins in the depressions. Such effects have been found to be of great significance in oil-field geology.[1]

The compaction of sediments by compression beneath a thick mass of reef limestone develops a basin in the strata at the base of the reef, as described by Bather.[2]

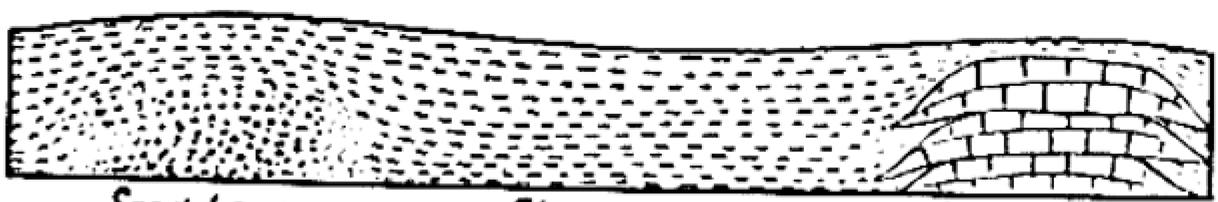


FIG. 7.—DIAGRAMMATIC CROSS-SECTION ILLUSTRATING THE DEVELOPMENT OF DOMES OVER A SAND LENS AND A CORAL REEF, DUE TO THE GREATER COMPACTION OF THE CLAYS. SCALE VARIABLE

Effects of Moving Ice.—Weak rocks beneath and in front of glaciers, grounded floe-ice, or icebergs, are brecciated, faulted, and folded by the drag of the moving ice. Such disturbances are characteristic of glacial deposits, but contortion can also be produced in older strata over which the ice has dragged, if these are sufficiently yielding. Sigmoidally curved thrust planes, as well as complex folds indicative of high plasticity, are associated in glacially disturbed sediments, as also in the stratified till deposited by retreating glaciers (see Fig. 8). In the latter, however, these structures do not result solely from deformation by the ice, but are in part inherited from the englacial debris, in which they were already present.[3] The folding of Tertiary

[1] Mehl, M. G., 'The Influence of Differential Compaction on the Attitude of Bedded Rock': *Science*, Vol. 1, 1920, p. 520. Powers, S., 'Reflected Buried Hills and their Importance in Petroleum Geology': *Econ. Geol.*, Vol. 17, 1922, pp. 233–59. See also the section on supratenuous folding, pp. 85–86.

[2] Bather, F. A., 'Reef Structures in Gotland': *Proc. Geol. Assoc. Lond.*, Vol. 25, 1914, pp. 225–8.

[3] For further information concerning glacial tectonics, the following works should be consulted: Slater, G., 'Studies in Glacial Tectonics': *Proc. Geol.*

brown coals in Germany by Pleistocene ice-sheets affords an excellent example of glacial tectonics (Fig. 49E).[1]

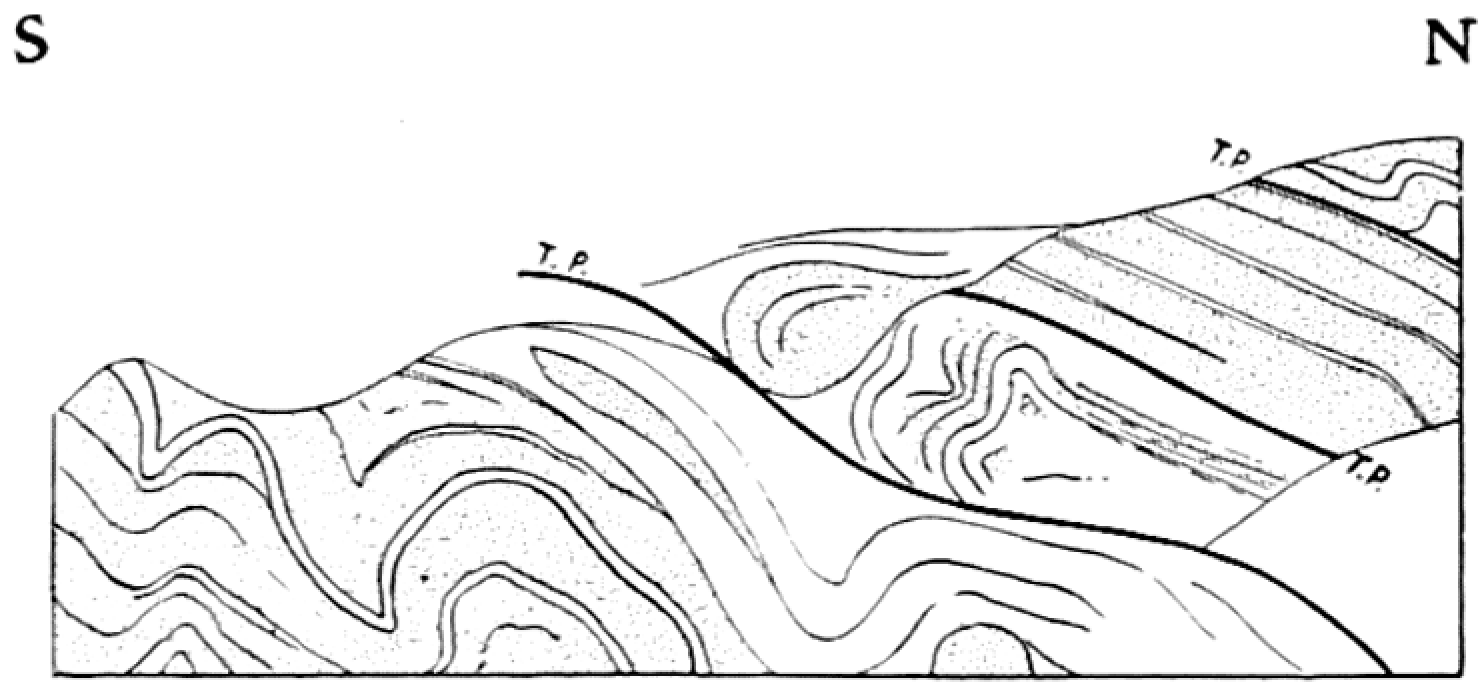


FIG. 8.—GLACIALLY DISTURBED UPPER CRETACEOUS SANDS AND CLAYS IN THE MUD BUTTES OF ALBERTA

(After Slater, 1927)

The ice moved from north to south. Length of section, 160 ft.; T.P., thrust planes.

Subaqueous Slumping, Sliding, or Gliding.—When sediments are deposited on a surface sloping near their angle of rest, they are liable to become unstable, and to slip off the floor on which they were deposited. This can result from the removal of support, as for example by the reduction of water level of a lake in which the sediments were deposited, or from earth tremors, or from instability due to the added weight of newly deposited sediments. Subaqueous slumping frequently occurs on slopes greater than 10°–15°, but it may also take place where the slope is as low as 2½°.[2] It is common on submarine fault scarps, on the continental slope in active regions of the crust, on the margins of subaqueous scour channels, and in deltas.

A slide may develop along the surface of unconformity at the base of an unstable sedimentary series, but more often it takes

Assoc. Lond., Vol. 37, 1926, pp. 392–400. *Ibid.*, Vol. 38, 1927, pp. 157–216. 'Structure of the Mud Buttes and Tit Hills in Alberta': *Bull. Geol. Soc. Amer.*, Vol. 38, 1927, pp. 721–30.

[1] Klein, G., *Die Deutsche Braunkohle-Industrie*: Bd. 1, Tl. 1, Halle, 3rd edn., 1927 (see p. 78 *seq.*).

[2] Grabau, *Principles of Stratigraphy*: 1st edn., 1913, pp. 780–5. Twenhofel, W. H., *Treatise on Sedimentation*: 2nd edn., 1932, pp. 739–44.

place along a stratification plane within the deposit. The most clearly recognizable structural result is complex deformation of the slumped mass, unconsolidated water-saturated beds and partially consolidated material yielding in different ways. The unconsolidated superficial sediments are highly plastic, and are thrown into involved fold structures, while the underlying beds which have been partly consolidated show less complex folding, minor thrusting, and, in some cases, brecciation. In the Tertiary rocks of Peru and Ecuador, thousands of feet of sediments have been affected by slipping, which took place mainly along low angle normal faults. Displacements of the order of 2 or 3 miles are believed to have occurred along these faults, with the development of breccias, complex minor folds, and other structures.[1]

The contortion of stratification planes into complex folds caused by gliding is termed *slip bedding* (see Fig. 9).

When a slumped mass slides down on to undisturbed sediments it may later be covered by younger deposits. Severely disturbed beds will then be found between undisturbed strata, an arrangement that is known as *intraformational contortion* or *corrugation*. Many examples of this are known in the geological column. In the Ordovician rocks of Girvan and the Kimmeridgian of East Sutherland, for example, remarkable breccias associated with strata showing slip bedding occur. Detailed investigations of these rocks suggest that submarine disturbances associated with faulting caused subaqueous slides, in which blocks of consolidated rocks were involved, giving rise to the breccias.[2]

Miller,[3] however, has criticized the interpretation of other

[1] Baldry, R. A., 'Slip-Planes and Breccia Zones in the Tertiary Rocks of Peru': *Quart. Journ. Geol. Soc.*, Vol. 94, 1938, pp. 347–58. Brown, B. B., 'On a Theory of Gravitational Sliding Applied to the Tertiary of Ancon, Ecuador': *ibid.*, pp. 359–70.

[2] Henderson, S. M. K., 'Ordovician Submarine Disturbances in the Girvan District': *Trans. Roy. Soc. Edinburgh*, Vol. 58, Pt. 2, 1935, pp. 487–509. Bailey, E. B., and J. Weir, 'Submarine Faulting in Kimmeridgian Times; East Sutherland': *Trans. Roy. Soc. Edinburgh*, Vol. 57, Pt. 2, 429–67.

[3] Miller, W. J., 'Intraformational Corrugated Rocks': *Journ. Geol.*, Vol. 30, 1922, pp. 587–610 (with bibliography).

examples of intraformational contortion as being due to slumping, and deformation by moving ice, the melting of buried ice blocks, flowage under gravity, the hydration and swelling of clays, and tectonic deformation may be mentioned among possible additional causes. It will therefore be realized that

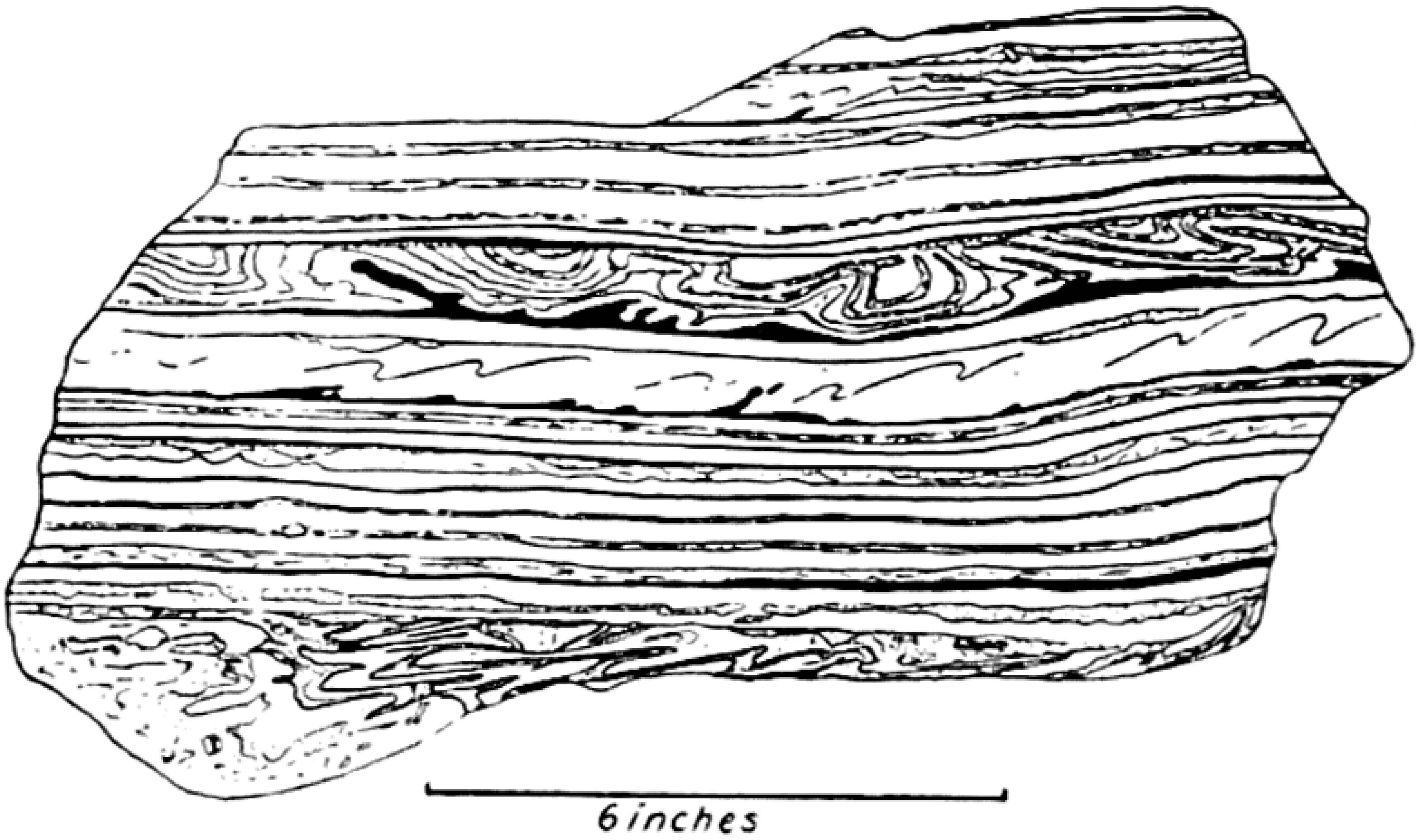


FIG. 9.—SLIP BEDDING IN A SPECIMEN OF ORDOVICIAN SHALE FROM THE GIRVAN DISTRICT

(After Henderson, 1935)

the recognition of slump-structures is not easy, especially in highly disturbed rocks, and the criteria for it must be carefully evaluated in the field.[1]

[1] See Boswell, P. G. H., *The Middle Silurian Rocks of North Wales*: London, 1949; 'The Tectonic Problems of an Area of Salopian Rocks in North-Western Denbighshire': *Quart. Journ. Geol. Soc.*, Vol. 93, 1937, pp. 284–321; Jones, O. T., 'On the Sliding or Slumping of Submarine Sediments in Denbighshire, North Wales, during the Ludlow Period': *ibid.*, pp. 241–83; 'The Geology of the Colwyn Bay District: A Study of Submarine Slumping during the Salopian Period': *Quart. Journ. Geol. Soc.*, Vol. 95, 1940, pp. 335–82. An excellent bibliography of soft-rock disturbances is given by Beets, C., 'Miocene Submarine Disturbances of Strata in Northern Italy': *Journ. Geol.*, Vol. 54, 1946, pp. 229–45. See also Fairbridge, R. W., 'Submarine Slumping and Location of Oil Bodies': *Bull. Amer. Assoc. Petrol. Geol.*, Vol. 30, 1946, pp. 84–92; Rich, J. L., '. . . Criteria for Recognition of Slope Deposits . . .': *ibid.*, Vol. 34, 1950, pp. 717–41.

The somewhat similar structures, produced as a result of chemical changes involving an increase or decrease in the volume of rocks such as salt deposits and limestones, are termed *enterolithic*.[1]

In connexion with the development of penecontemporaneous structures and slumping, the possible influence of the thixotropy of sediments has been stressed.[2] A thixotropic mixture of sediment with water changes from the gel condition to a fluid (sol) on agitation or shock, and reverts to a gel on standing. In mixed sedimentary successions some beds may be in a thixotropic condition and others not, a condition that might account for the brecciation of certain strata in slump-masses.

Gravity-Collapse Structures.—In south-west Persia, structures in Mesozoic and Tertiary sediments folded during Tertiary orogenesis, have been explained by postulating the collapse under gravity of great sheets of limestone after the removal of their supporting cover. These have been termed *gravity-collapse structures*[3] (see Fig. 10). When the crest of a sinusoidal fold in interbedded limestones and shales is broken through by erosion, and the limestone sheets move apart by slipping downwards and outwards over underlying shales, a *knee fold* is first formed in the limestones. If this is accentuated, the resulting structure resembles the side of a house, with its sloping roof and vertical wall, and is therefore called a *roof and wall structure.* A *cascade* is a group of subsidiary folds developed in limestone by collapse down a dip slope, and a *slip sheet* is one that has fractured at its base and slipped down a dip slope. A *flap* is part of a limestone sheet that has bent over backwards, away from the fold axis, without fracturing. It lies with the lower

[1] Grabau, A. W., *Principles of Stratigraphy*, 1st edn., 1913, p. 756 *et seq.*

[2] Boswell, P. G. H., 'The Thixotropy of Certain Sedimentary Rocks': *Science Progress*, No. 143, 1948, pp. 412–22.

[3] Harrison, J. V., and N. L. Falcon, 'Collapse Structures': *Geol. Mag.*, Vol. 71, 1934, pp. 529–39; 'Gravity Collapse Structures and Mountain Ranges, as exemplified in South-western Persia': *Quart. Journ. Geol. Soc.*, Vol. 92, 1936, pp. 91–102. Lees, G. M., 'The Geology of the Oilfield Belt of Iran and Iraq': in *The Science of Petroleum*: Oxford, 1938, pp. 140–8.

beds uppermost, on the younger strata exposed in the valley adjacent to the arched structure.

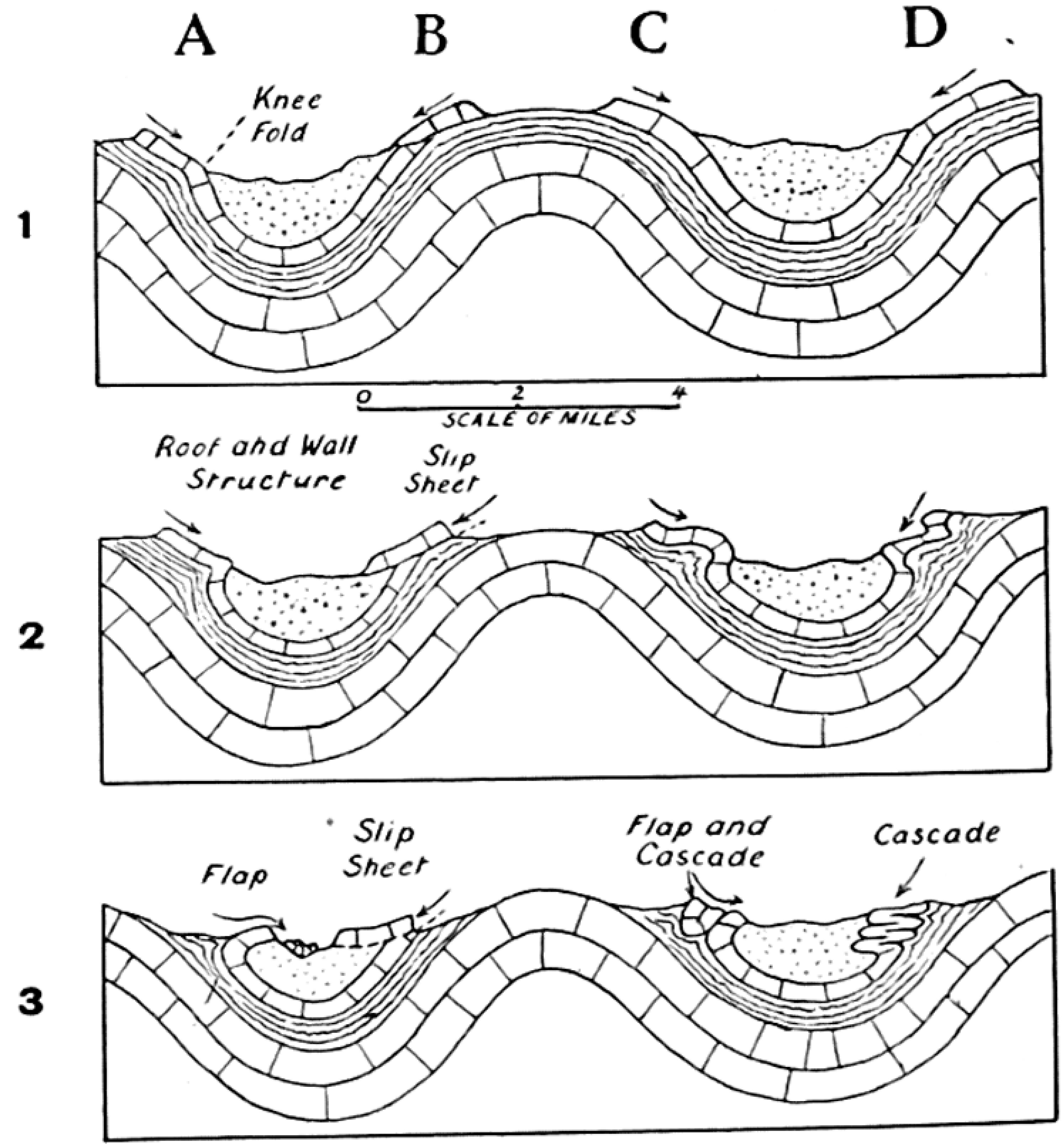


FIG. 10.—GRAVITY COLLAPSE STRUCTURES

(After Harrison and Falcon, 1934)

The diagrams 1, 2, and 3 represent successive stages in the development of a *flap* (sequence below A), a *slip sheet* (below B), a *flap* and *cascade* (below C), and a *cascade* (below D). The arrows indicate the direction in which the limestone sheets have slipped over the shales, which are shown by stippling. Vertical and horizontal scales the same.

Rock-flowage under Gravity.—Gravity-collapse structures, although local and superficial, serve to demonstrate that complex structures may develop on a large scale by the plastic flow of rocks down the flanks of elevations and into troughs.

Bain and de Terra have also demonstrated the folding of solid rocks by gravitational sliding,[1] but the process is given much wider significance by some authors, who attribute the folding and faulting of orogenic belts to flowage either in the stage of geosynclinal deposition, or later, after the upraising of the sediments by vertically-acting forces. Further reference to these topics is made in Chapter III.

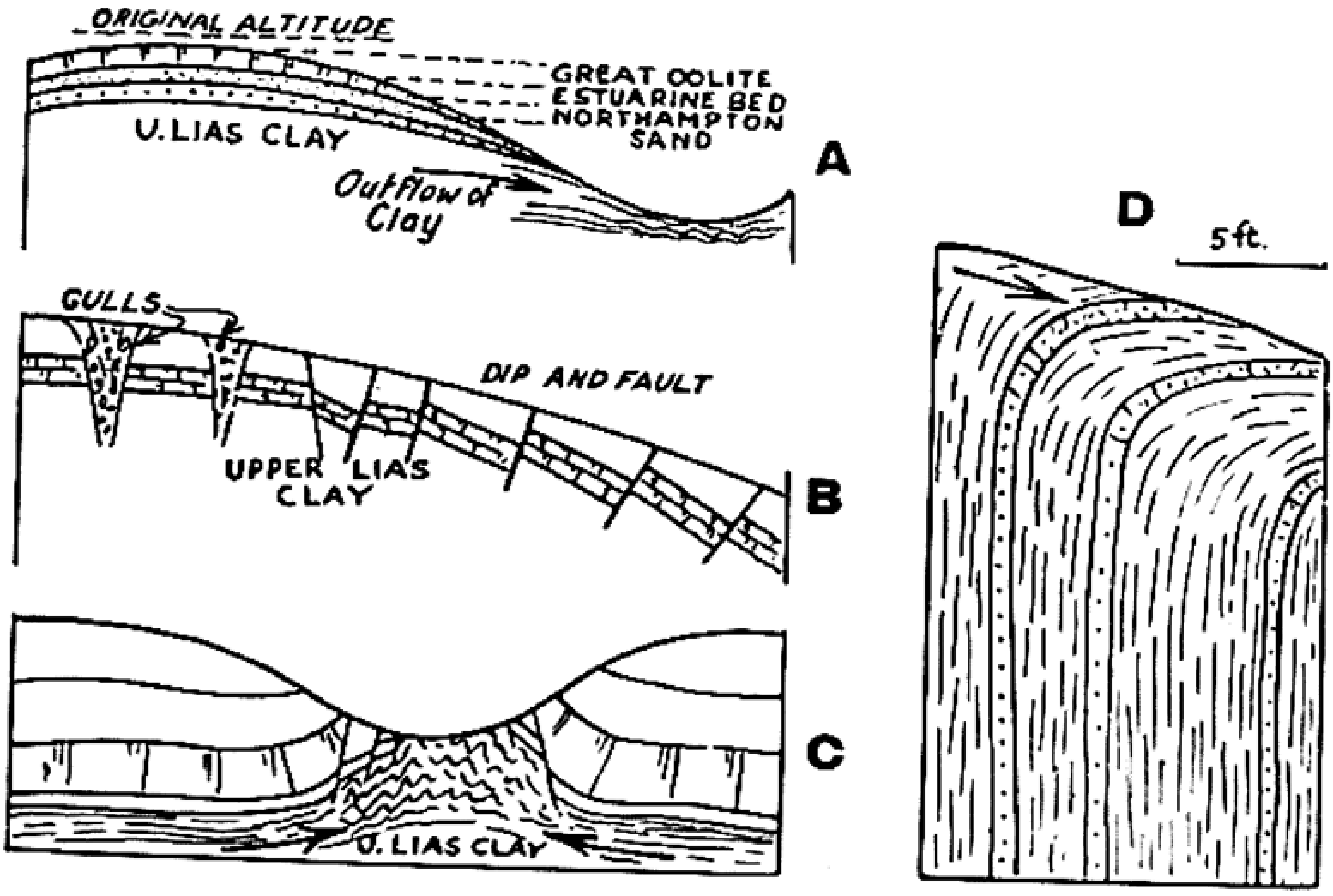


FIG. 11.—SUPERFICIAL STRUCTURES

A, B, C, Large-scale superficial structures in Northamptonshire (after Hollingworth and others).

A, 'Camber' in beds overlying the Upper Lias Clay; B, Gulls, and 'dip-and-fault'; C, Valley bulge due to rise of clay beneath river valley.

D, Hill-creep in vertical sandstones and shales, Studley Park, Melbourne.

Structures developed by the breaking away of cliffs, and in landslides and mine subsidences, may also be mentioned in connexion with collapse under gravity.[2]

[1] Bain, G. W., 'Flowage Folding': *Amer. Journ. Sci.*, Vol. 22, 1931, pp. 503–30. de Terra, H., 'Structural Features in Gliding Strata': *ibid.*, Vol. 21, 1931, pp. 204–13.

[2] Sharpe, C. F. S., *Landslides and Related Phenomena*: Columbia University Press, 1938. Heim, A., *Über Bergstürze*: Zürich, 1882. Bendel, L., *Ingenieurgeologie*, Teil II: Vienna, 1948.

The geological significance of superficial structures is emphasized by Hollingworth and others in work on the Northampton Ironstone Field,[1] where clays underlie ironstone, limestone, and other rocks. The harder beds, as a result of subsurface erosion and valley-ward outflow of the clay, develop various structures termed *camber*, *gulls*, and *dip-and-fault*, and the squeezing up of clay in the valley flows where the superincumbent rocks have been removed, produces anticlines called *valley bulges* (Fig. 11). Such structures must be much more common than has hitherto been realized, for evidence of the flow of soil and weathered rock down hill slopes (*hill-creep*) is ubiquitous. Hill-creep modifies the dip in the superficial zone, which in soft rocks may extend down for many feet. Surface dip-readings are useless in rocks affected by creep (Fig. 11D).

Salt Domes.—Where a bed of rock salt is buried beneath later sediments, the salt tends to rise because of its low specific gravity, and if the overlying beds are sufficiently plastic the salt intrudes them as plugs or *salt domes*. Complex folds in the salt are the result of flowage, and doming of the sediments over the intrusive mass, with drag effects along its margin, are also formed. The forces involved are hydrostatic, and salt domes and related features may therefore form independently of diastrophism (see pp. 94–6 for further details). In areas affected by diastrophism, salt-tectonics may be very complex.

[1] Hollingworth, S. E., J. H. Taylor and G. A. Kellaway, 'Large-scale Superficial Structures in the Northampton Ironstone Field': *Quart. Journ. Geol. Soc.*, Vol. 100, 1944, pp. 1–44. Hollingworth, S. E., and J. H. Taylor, 'An Outline of the Geology of the Kettering District': *Proc. Geol. Assoc.*, Vol. 57, 1946, pp. 204–33.

the clay, which have moved relatively to each other along the planes separating them. These planes are termed *slip planes* or *shearing planes*; they occur in two vertical sets, approximately at right angles to each other and at 45° to the directions of elongation and compression (see Fig. 17). The intersection of the

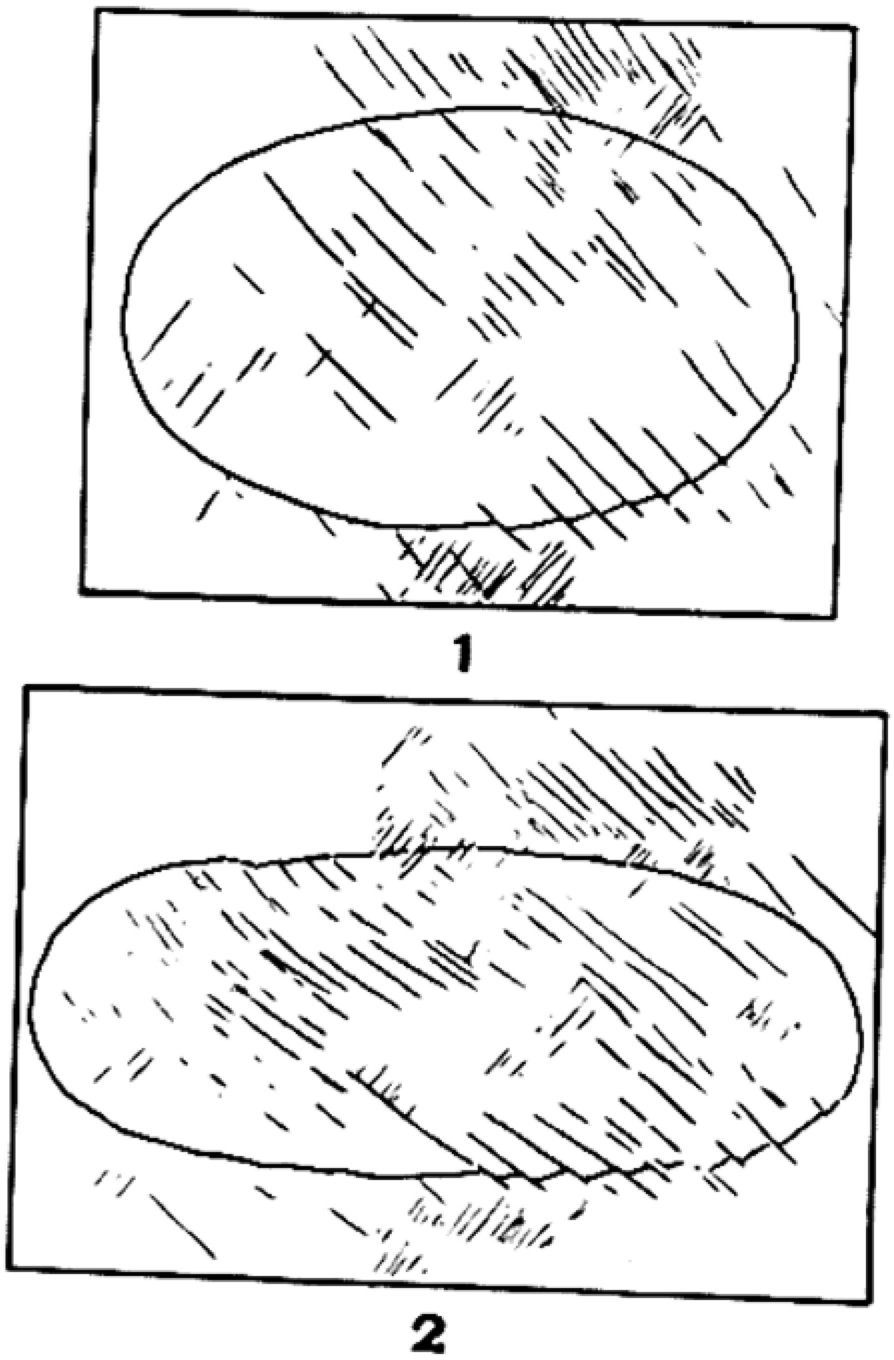


FIG. 17.—TWO STAGES IN THE DEFORMATION OF A CIRCLE INSCRIBED ON CLAY, SUBJECTED TO PURE SHEAR

(Drawn from photos by Cloos, 1930)

Note the two intersecting sets of shearing planes.

planes, therefore, corresponds with the mean strain axis, and this relationship, which is of general application, is very useful in the interpretation of many geological structures.

In crystalline materials such as mild steel, the relationship between slip planes and grain deformation may be observed. Polished steel test pieces subjected to tension or compression beyond the elastic limit and examined by special methods of

illumination, reveal a pattern of dark and light bands called *Lüders' lines* or *slip bands*.[1] Like the shearing planes in clay, these slip bands intersect in the mean strain axis if a triaxial strain ellipsoid is developed, and the angles between them are bisected by the greatest and least strain axes, with which they make an angle of approximately 45°. They represent zones of

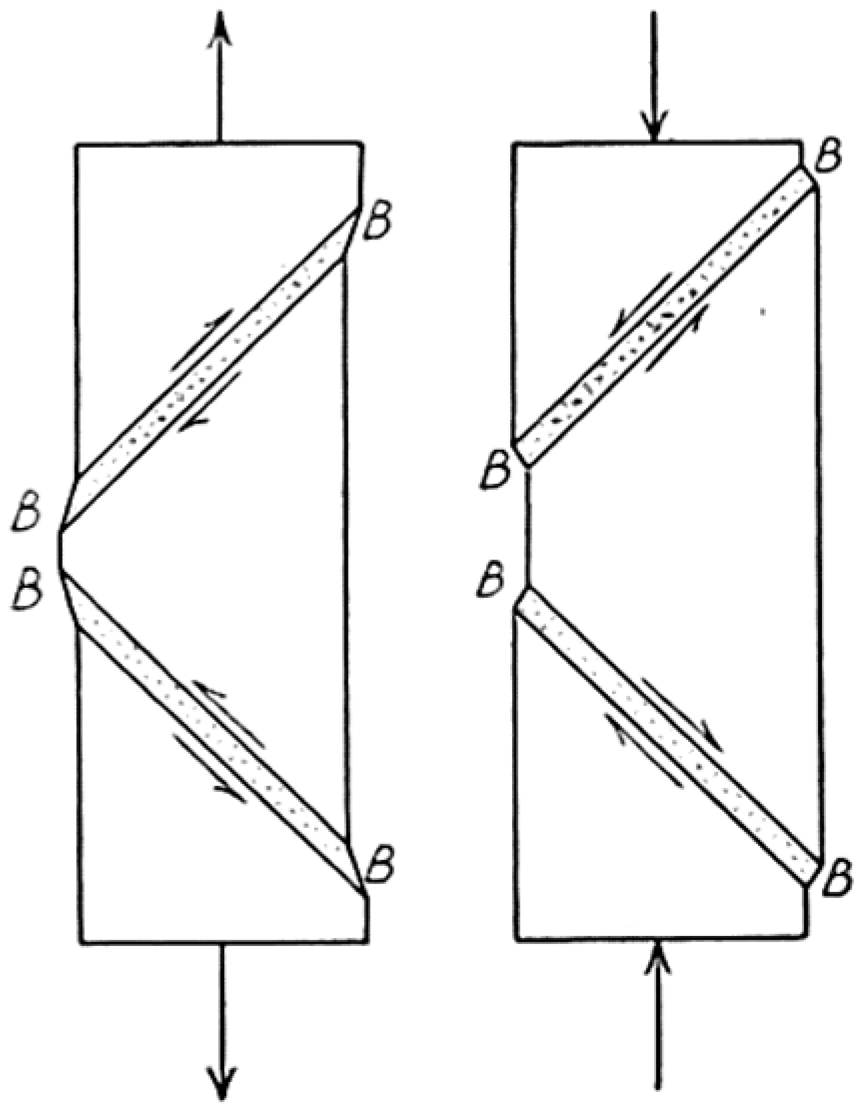


FIG. 18.—FORMATION OF STRAIN FIGURES BY TENSION AND BY COMPRESSION

(After Nádai, *Plasticity*)

B–B, slip bands, the arrows indicating the direction of movement along the bands. The appearance of light and dark bands on the surface is due to the variable angle of reflection of incident light.

maximum grain deformation in the steel, and lie between less strongly deformed parts which have slipped relatively to each other parallel to the slip bands (see Fig. 18). Under the microscope, it may be seen (Fig. 19) that the twin and glide planes in the strongly deformed grains of the slip bands show a marked

[1] Nádai, A., *Plasticity*: New York, 1931, Chap. 16. Turner, T. H. and J. D. Jevons, 'The Detection of Strain in Mild Steels': *Journ. Iron and Steel Inst.*, Vol. 3, 1925, pp. 169–89 (with many references).

parallel orientation.[1] Such preferred orientations of the crystallographic structure elements in deformed rocks have been the subject of detailed study (see Chap. VII).

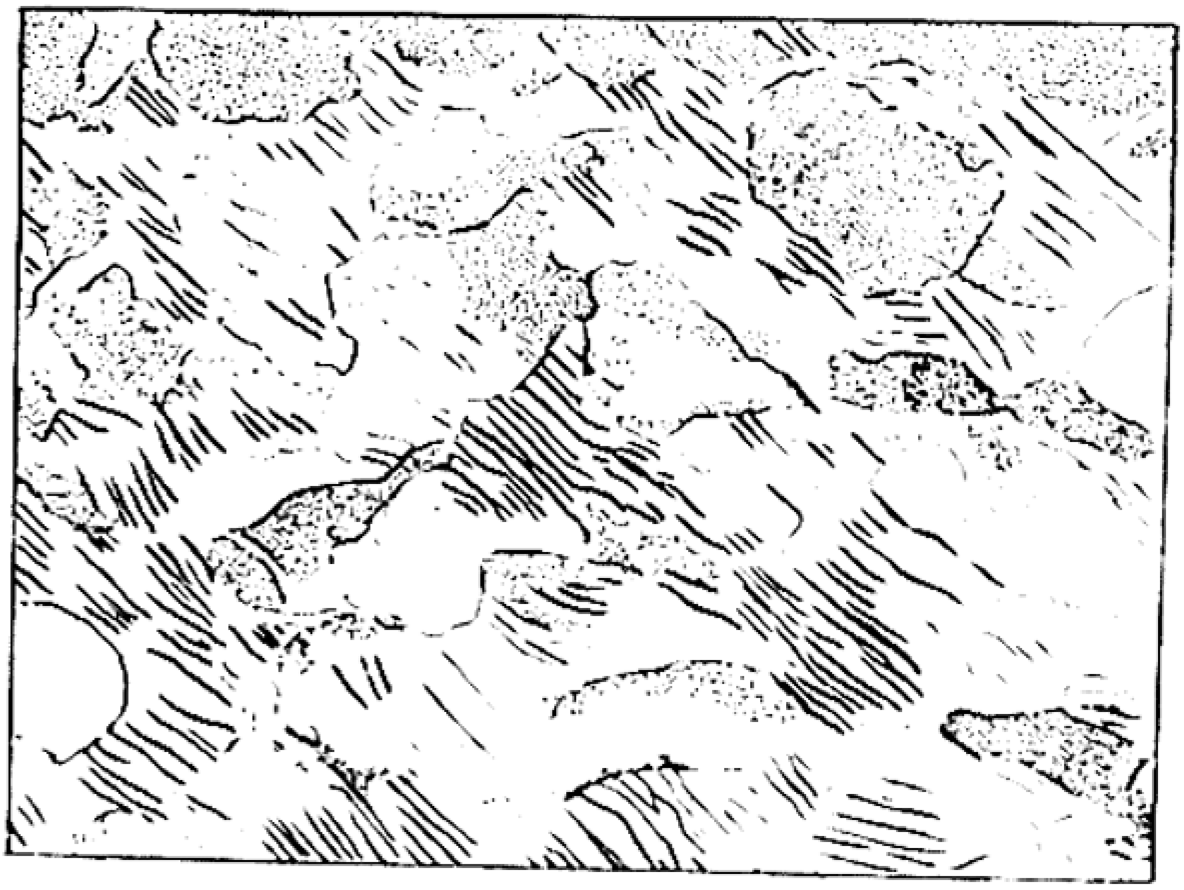


FIG. 19.—MICROPHOTOGRAPH OF THE BORDER OF A LÜDERS' LINE DEVELOPED ON A POLISHED SPECIMEN OF MILD STEEL JUST AFTER THE YIELD POINT HAS BEEN REACHED

(After Nádai, *Plasticity*)

Note the tendency for the slip bands, which mark the intersection of glide planes in the individual crystals with the surface of the specimen, to be orientated parallel to a certain direction. Note also that grain rotation has occurred, some grains now appearing dark and others light under incident light.

Fracture.—Since the movements of the different parts of a body during plastic flow are localized along the slip planes, it is natural to expect that the ultimate rupture will also take place along these planes. This is actually found to be the case with many ductile materials, but fracture may also take place by pulling apart of the material in planes at right-angles to the axis of maximum strain. Brittle substances in tension characteristically fail in this manner, and in the experiment with clay described above, fracture may be induced by flooding the surface of the clay with water, which reduces its plasticity and

[1] Griggs has noted a similar orientation of the twin planes in experimentally deformed marble. See *Journ. Geol.*, Vol. 44, 1936, pp. 541–77.

causes *tension gashes* to open in the plane of the mean and least strain axes.

In compression, brittle substances fracture along planes that have a similar attitude relative to the principal strain axes as do the shearing planes in plastic materials. We have noted that the angle between the shearing planes in plastic clay and in steel is approximately 90°. To be more specific, it is found that the angle that is bisected by the axis of least strain (which corresponds with the axis of maximum stress in irrotational

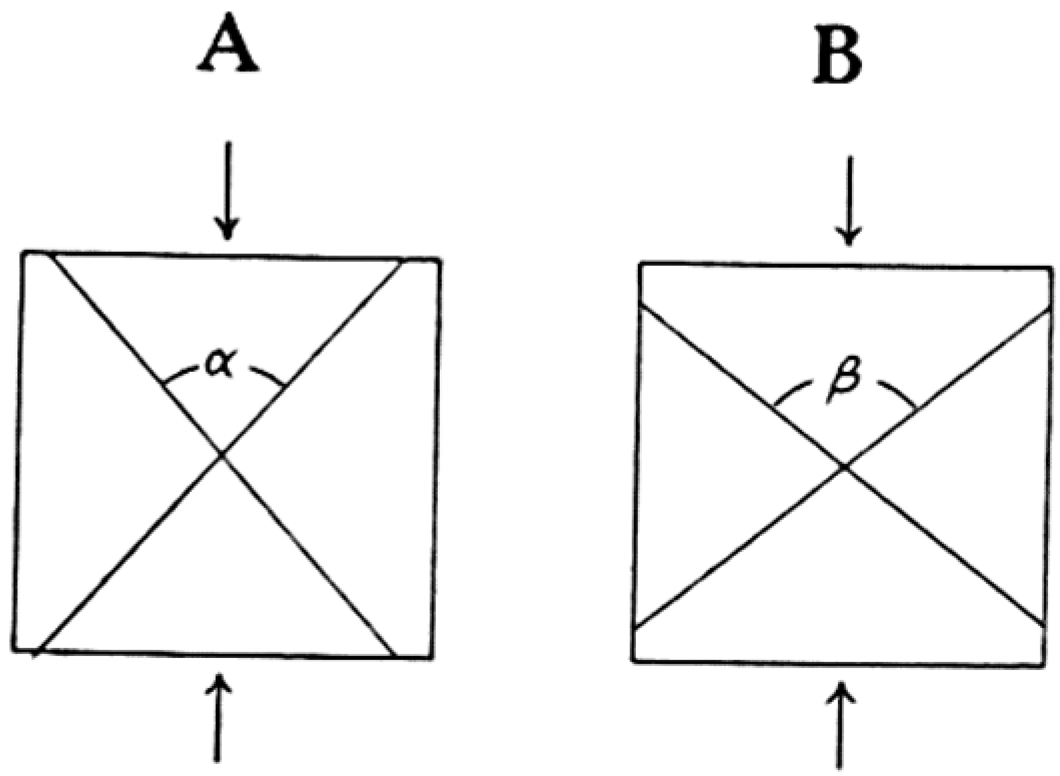


FIG. 20.—SHEARING PLANES IN BRITTLE AND DUCTILE SUBSTANCES SUBJECTED TO COMPRESSION

A, brittle substance: the angle (α) between the shearing planes enclose ing the axis of maximum stress (indicated by the arrows) is acute. B, ductil- substance: the angle (β) is obtuse.

strains) is acute for brittle substances and obtuse for ductile substances (Fig. 20). This relationship, the theoretical interpretation of which we need not consider here,[1] is of fundamental importance for the geologist, for as we have seen, rocks which are brittle at ordinary temperatures and pressures may yet be highly plastic when buried deep in the crust.[2] The same rock type observed at different localities may therefore have

[1] A useful summary of the various theories of strength and plastic yielding is given by Nádai (*Plasticity*, Chap. 12).

[2] Bucher, W. H., 'The Mechanical Interpretation of Joints': *Journ. Geol.*, Vol. 28, 1920, pp. 707–30; *ibid.*, Vol. 29, 1921, pp. 1–28.

a different angle of shear at each, this depending upon the physical conditions to which it was subjected when deformed.

Shearing Planes in Rotational Strains.—Since in rotational strains the position of the principal strain axes changes within the stressed body, the attitude of the planes along which slip should occur will also change. Once shearing planes are established, however, we are no longer dealing with a homogeneous body, for these planes then constitute pre-determined

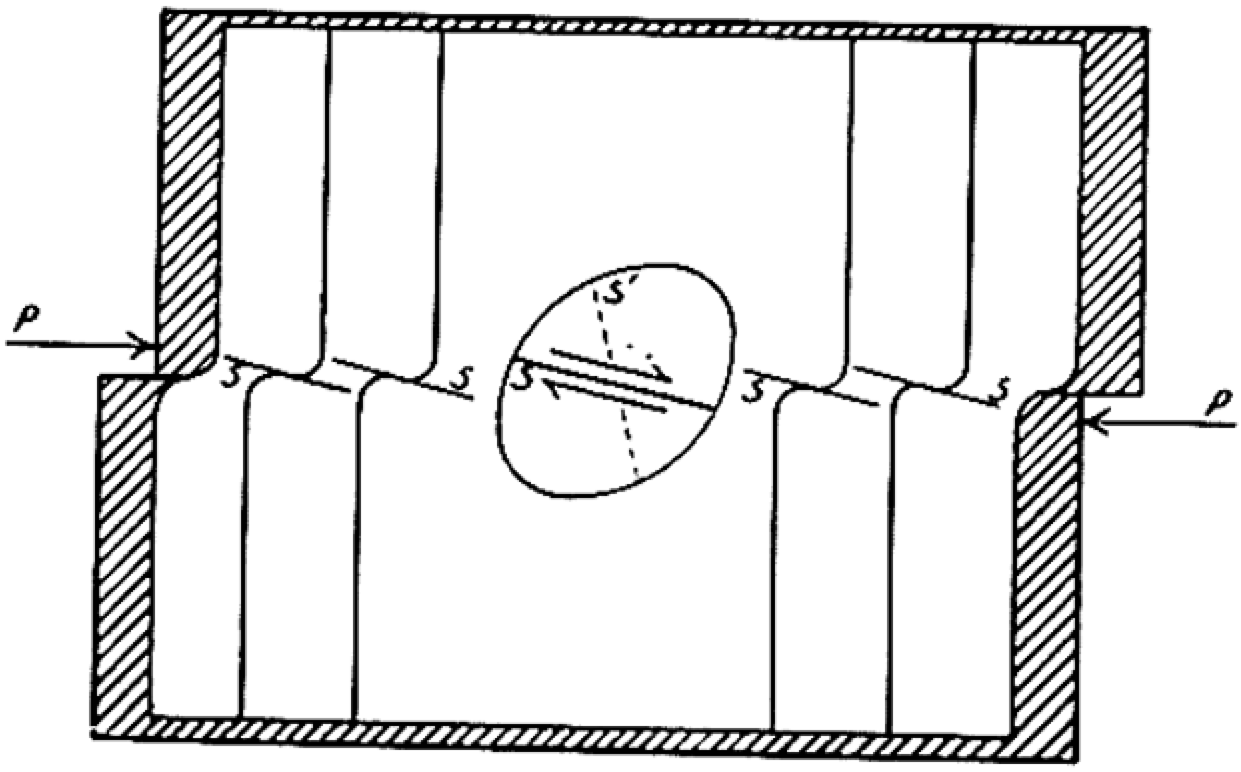


FIG. 21.—SHEARING PLANES (S) IN A CAKE OF CLAY RESTING ON TWO BOARDS THAT HAVE BEEN PUSHED ALONG THEIR CONTACT IN THE DIRECTION OF THE ARROWS P

(Based on Riedel, 1929)

An imaginary strain ellipse in the zone of deformed clay is shown (enlarged for convenience). Only one of the two possible directions of shearing S and S′ is developed—that which lies more nearly parallel to the direction of shearing movement between the boards.

directions of ease of slip for later deformation. Thus, unless the rotation of the theoretical shearing directions away from the planes already initiated is considerable, slip will tend to be restricted to the latter. Rotational strains, too, are brought about by shearing couples, and one of the sets of shearing planes lies closer to the direction of the couple than the other (see Fig. 21). Slip along these planes is easier than along the others, because the normal stresses acting on them are less. Also, the development of long shearing planes at an angle to

those along which the maximum movement is occurring is prevented by the slip along these latter planes, for this continually interferes with the slip along the other set. Thus we find that in finite rotational strains one set of shearing planes is better developed than the other.

Bending.—One of the most important types of heterogeneous strains in the deformation of rocks is bending. On the outer side of a bent plate, the stress is tensile, on the inner side, compressional, and in an intermediate position there is a surface neither shortened nor elongated, which is termed the *neutral surface*[1] (Fig. 22). In thick brittle strata the circumferential

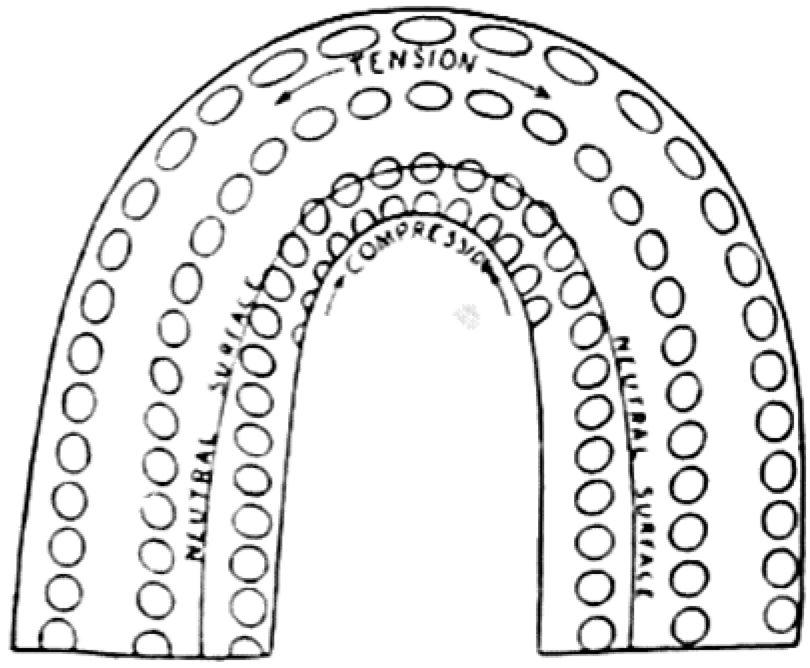


FIG. 22.—BENDING OF A BAR OF PLASTIC MATERIAL

(Adapted from Sander, *Gefügekunde der Gesteine*)

Circles inscribed before deformation indicate the state of strain in the various parts of the bar.

tension often causes radially arranged tension gashes or joints, which are usually best seen in anticlines. The effects of the compression in the cores of the folds are not often noticeable, but minor puckers sometimes arise in this position[2] (Fig. 23). Most rocks yield more readily in tension than in compression, so that the tension gashes form before the rock yields to the compressive stresses in the cores of folds. As the gashes form, the effective outer surface of the folded bed is shifted to the inner ends of the gashes, so that the neutral surface is moved

[1] Ickes, E. L., 'Similar, Parallel, and Neutral Surface Types of Folding': *Econ. Geol.*, Vol. 23, 1923, pp. 575–91.

[2] Sander, B., *Gefügekunde der Gesteine*: Vienna, 1930, p. 248.

closer to the core of the fold, and the cracks extend themselves, until they ultimately cross the whole stratum. In this way the effects of compression in the cores of folds in thick brittle beds is reduced.

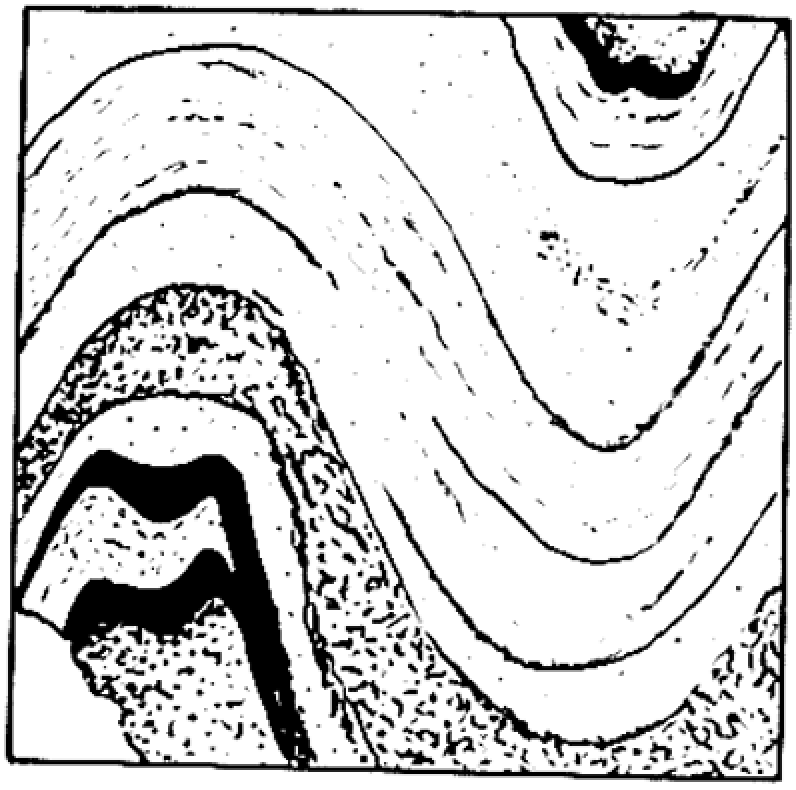


FIG. 23.—SUBSIDIARY BUCKLING IN THE CORES OF FOLDS IN QUARTZ PHYLLITE FROM INNSBRUCK, TIROL. $1\frac{1}{2}$ APPROX. (Diagrammatic, after Sander, *Gefügekunde der Gesteine*)

Torsion.—A body is in torsion when it is subjected to two force couples acting in parallel planes about the same axis of rotation, but in opposite senses (Fig. 24). Since Daubrée's classical experiments on the torsion of thin glass plates, torsion has often been referred to in text-books as the chief cause of joint sets intersecting at approximately 90°. Bucher, in criticizing this interpretation of the experiment,[1] points out that the complementary sets of cracks in the glass arise because of the thinness of the plates and the brittleness of glass, which allow the cracks developed on one side to pass through to the other. The axes of maximum strain on either side of the plate lie at an angle to each other, so that the major cracks, which are tension gashes, appear as intersecting sets on both sides of the plate. Comparable conditions may, however, exist in brittle rocks, such as coal, occurring in relatively thin sheets.

Plasticity and Flow of Rocks.—In interpreting the conditions under which observed rock structures were developed it is

[1] See references on p. 99.

essential to understand the physical state and properties of the rocks at the time of their deformation, which may range from fluid to rigid solid. While this is obvious in the case of igneous rocks, the conditions with sediments are also complex, since the properties of such rocks are profoundly affected by the changes they undergo in different environments.

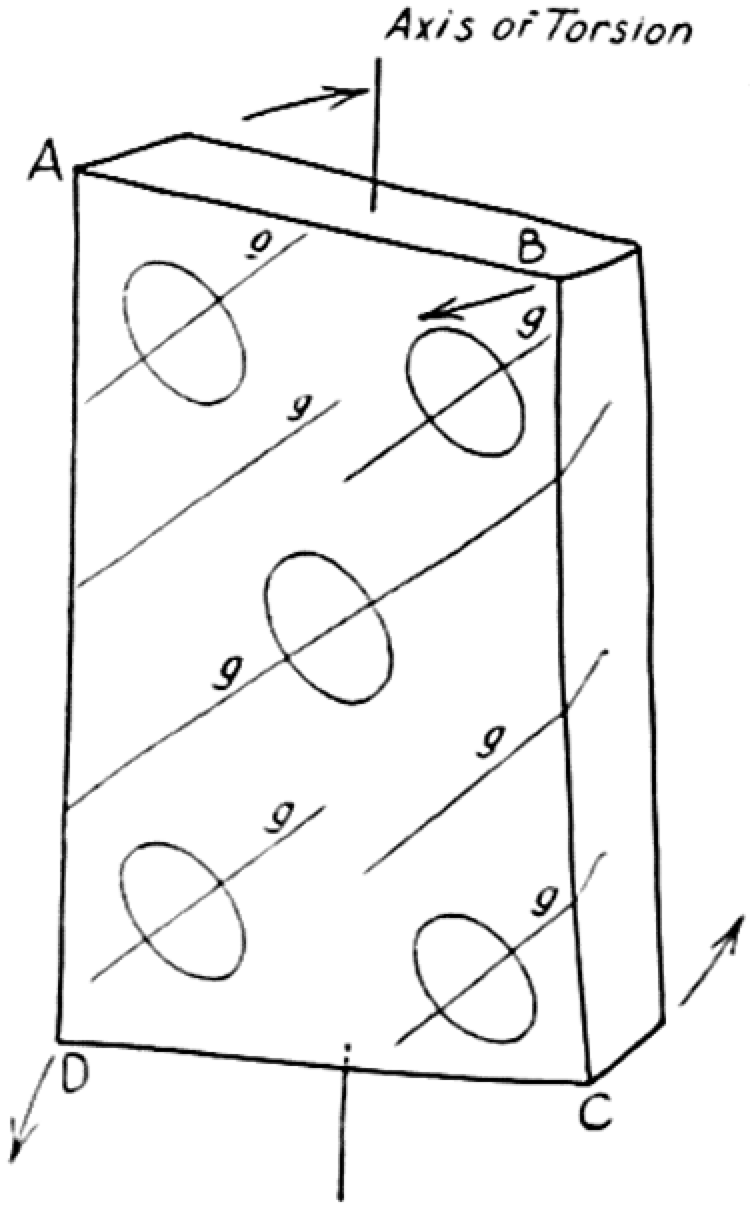


FIG. 24.—TORSION OF A PLATE

Circles drawn before deformation on the face ABCD are changed into ellipses as shown. Tension gashes *g* develop on the face ABCD. On the face opposite this, the ellipses are drawn out at right angles to those shown, and the tension gashes develop at right angles to *g*.

A sedimentary rock in which the grains are cemented by secondarily introduced bonding material such as silica, carbonates, or iron oxides may be brittle, whereas the same rock, if uncemented, may be plastic. The presence of pore-water in sedimentary rocks in which the grains are not cemented results in *hydroplasticity* as in wet clay.[1] On dehydration, clay becomes

[1] Term introduced in Shrock, R. R., *Sequence in Layered Rocks*: New York 1949, p. 152.

less plastic. The strength of loose granular materials such as sand is greatly reduced if the water content attains a certain value, at which pressure is transmitted largely through the pore-water rather than from grain to grain. Thus in normal sedimentary rocks containing sand and clay, structures may develop before diagenesis is complete, that is under soft-rock conditions, and subsequent lithification by cementation and other processes may then indurate the rocks so that, to make them plastic a second time, quite severe changes in the environment would be required. There is reason to believe that strong folding of many sedimentary successions took place before complete lithification[1] and thus under conditions in which the rocks must be regarded as more or less loose granular aggregates rather than as solid bodies. In the flowage of strongly bonded rocks, the individual crystal grains are involved and the effects of crystal-plasticity, visible under the microscope, afford information on the strain which the rock has undergone. This is more fully discussed in Chapter VII.

[1] Hills, E. S., 'The Silurian Rocks of the Studley Park District': *Proc. Roy. Soc. Vict.*, Vol. 53, 1941, pp. 167–91.

Chapter III

MAJOR CRUSTAL STRUCTURES

1. *THE TECTONIC FRAMEWORK OF THE CRUST*

OUR knowledge of the structure of the earth's crust is derived largely from the continents, which constitute one of the first-order tectonic elements of the earth, the other being the ocean basins. From late Pre-Cambrian time onwards there may be recognized in or marginal to the continents, two types of regions of markedly different geological history and tectonic significance—the *mobile belts*[1] and the resistant blocks or *cratons*. Geological mobility is attested by earthquakes, igneous activity, strong deformation of rocks, and elevation or depression, which the mobile belts exhibit in various degrees. Depressed zones that receive thicknesses of sediments very much greater than adjoining regions are *geosynclines*,[2] and elevated zones either within geosynclines or bordering them, *geanticlines*. Both are typically associated with mobile belts, but geosynclines, because of their thick sedimentary filling, are preserved in the geological column and bear witness to the existence of former mobile belts.

During Palaeozoic and later time, certain persistent zones of great mobility may be traced virtually continuously as belts of world-wide extent, in which after preliminary stages including sedimentation, volcanic activity and complex oscillations,

[1] Bucher, W. H., *The Deformation of the Earth's Crust*: Princeton, 1933.

[2] The word 'geosyncline' has been used in a wide variety of connotations, an excellent review of which is given by Glaessner, M.F. and C. Teichert, 'Geosynclines: A Fundamental Concept in Geology': *Amer. Journ. Sci.*, Vol. 245, 1947, pp. 465–82, 571–91.

folding, faulting and uplift gave rise to mountain chains. The uplift of such chains appears to take place chiefly by vertical movements after strong deformation of the geosynclinal sediments, and indeed the Palaeozoic compacted mobile belts

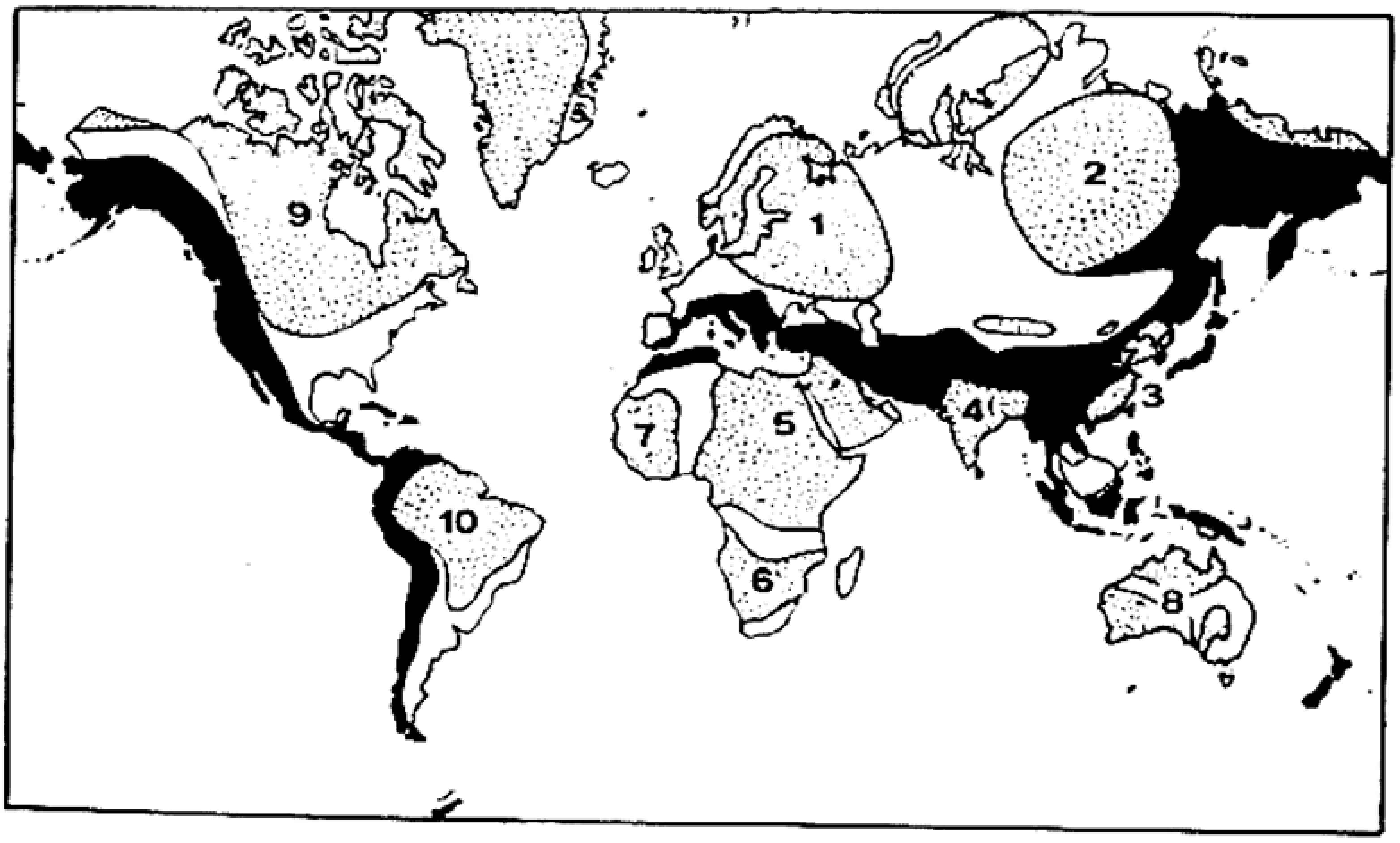


FIG. 25.—GENERALIZED MORPHO-TECTONIC MAP OF THE LAND AREAS OF THE GLOBE

A. Shields and tables, shown by stippling.
1, Russian table and Baltic shield; 2, Siberian shield (Angaraland); 3, Sinian tables; 4, Indian shield; 5, Ethiopian-Arabian table; 6, Karroo table; 7, West African shield; 8, Australian shields and blocks; 9, Canadian shield; 10, Brazilian shield.
B. Belts of Mesozoic and Tertiary orogeny, shown in solid black.
C. Belts of Palaeozoic orogeny, blank.

Note the general correspondence of existing mountain chains of Alpine type with the Mesozoic and Tertiary orogenic belts; of ranges of lesser elevation with the Palaeozoic orogenic belts, and of plain and plateau areas with the shields and tables.

are commonly rejuvenated by uplift associated with faulting, as in the Appalachians, the Urals, and the highlands of eastern Australia[1] (Fig. 26). Mobile belts of the above type are

[1] For the Appalachians see Keith, A., *Bull. Geol. Soc. Amer.*, Vol. 34, 1923, pp. 330–4; for the Urals, Kober, L., *Der Bau der Erde*: Berlin, 1928, pp. 230–2; for Australia, David, T. W. E., *Geology of the Commonwealth of Australia*: London, 1950.

orogens, with which the cratons may be sharply contrasted.[1] Cratons are the relatively stable blocks that comprise the major portion of continents, and perhaps also of the ocean basins, where belts of greater mobility such as the Mid-Atlantic Swell may also be recognized. The larger cratons have cores or nuclei of Pre-Cambrian rocks, which form broadly arched plateaux and are therefore known as *shields*, but these may be covered in places by little-disturbed younger rocks, thus constituting *tables* (see Fig. 25). Smaller areas of Pre-Cambrian rocks are termed *blocks*. The deformed rocks of older

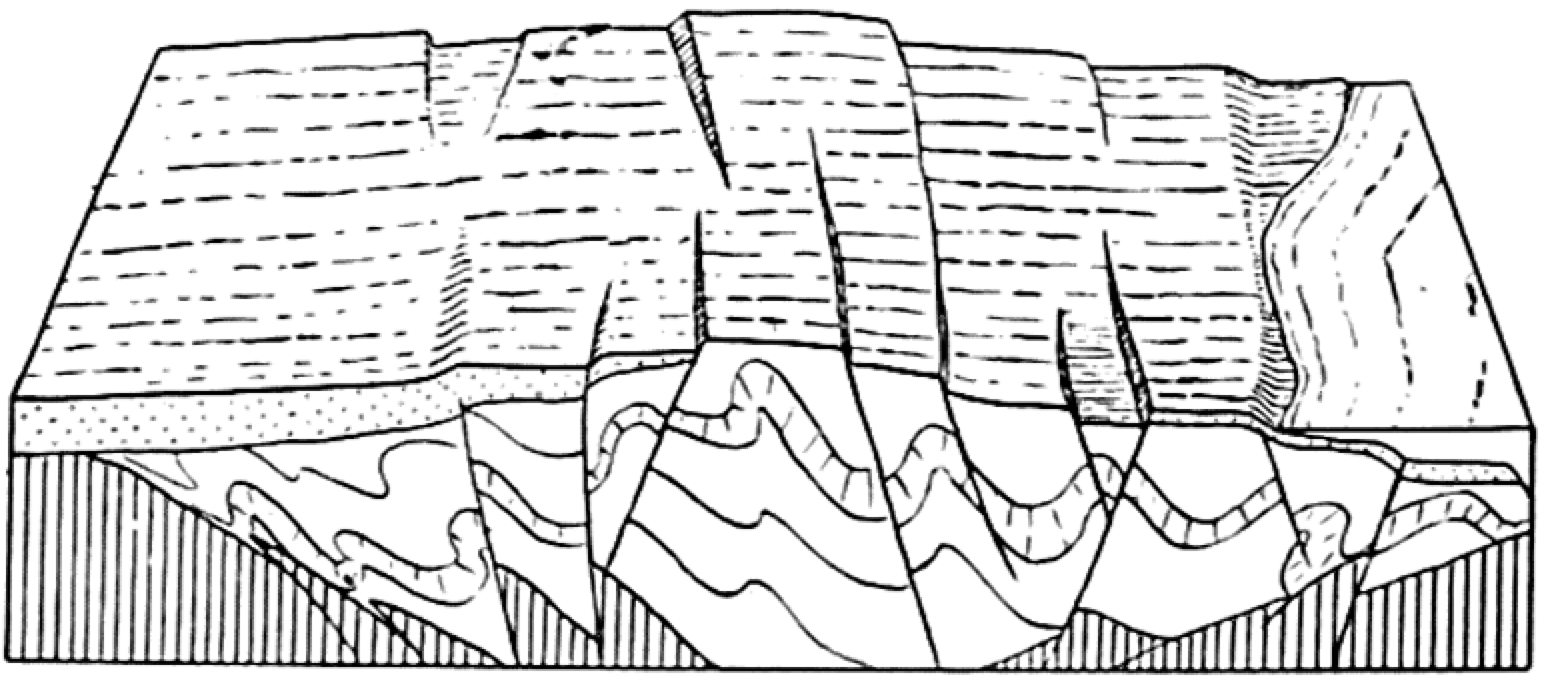


FIG. 26.—BLOCK DIAGRAM OF PORTION OF THE EASTERN AUSTRALIAN HIGHLANDS—A REJUVENATED PALAEOZOIC MOBILE BELT

Showing the crystalline basement, the folded rocks of the compacted mobile belt, and a partial cover of younger rocks, the whole being affected by Cainozoic warping and faulting.

orogens, buttressed with igneous intrusions, are welded to the nuclei and act as part of the cratonic regions for younger orogens.

Orogens and cratons are clearly recognizable in the Upper Proterozoic, and even in older Pre-Cambrian rocks there are local indications of similar features, but zones of world-wide extent are not known. Archean rocks are everywhere strongly folded and faulted, and the tectonic conditions accompanying

[1] The terms *resistant block* (or *mass*), *stable block*, or *rigid block* are approximately synonymous with craton. The term *orogen* is here used without the special significance given to it by Kober (see p. 60).

their deformation were, very likely, different from those of later periods.[1]

Earth-movements producing strong folding and reverse faulting indicative of compression have been termed *orogenic* movements, since such structures are found in many mountain chains, and the broad warping and vertical displacements of cratons are termed *epeirogenic* movements.[2] In some young high mountain chains, however, folding is subordinate to vertical uplift,[3] and the folding of orogenic belts is regarded by some geologists merely as a consequence of vertical movements either during or after the geosynclinal stage (see pp. 64–6). Again, fracturing, warping, folding, uplift, and depression are exhibited in all degrees of intensity and some epeirogenic-type movements in cratons are possibly causally connected with orogenic-type movements in mobile belts. The distinctions between the two types lie firstly in the wide regional effects of epeirogenesis and the narrow zonal effects of orogenesis; the great intensity of structural development in orogenesis, as against the less violent movements of epeirogenesis, and the predominance of vertical movements in epeirogenesis, and of lateral movements in orogenesis.

2. *TECTONIC PATTERNS*

Regular geometrical relationships between the structural features of broad regions or even of the whole earth have been recognized by several authors. Lothian Green's Tetrahedral Theory, propounded in 1875, still affords the only explanation

[1] Various suggestions concerning early Pre-Cambrian tectonics will be found in the following works: Wilson, J. T., 'The Origin of Continents and Precambrian History': *Trans. Roy. Soc. Canada*, Vol. 43, 1949, pp. 157–84; 'Some Major Structures of the Canadian Shield': *Canadian Min. & Met. Bull., Trans.* Vol. 52, 1949, pp. 231–42; Hills, E. S., 'Some Aspects of the Tectonics of Australia': *Proc. Roy. Soc. N.S.W.*, Vol. 79, 1946, pp. 67–91; Hills, G. F. S., *The Formation of Continents by Convection*: London, 1947.

[2] Gilbert, G. K., 'Lake Bonneville': *U.S. Geol. Surv.*, Mon. No. 1, 1890, p. 340.

[3] E.g. the Andes, and eastern New Guinea: see Picard, L., 'La Structure du NW. de l'Argentine etc.': *Bull. Soc. Geol. France*, Vol. 18, 1948. David, T. W. E., *Geology of the Commonwealth of Australia*: London, 1950.

ex hypothesi of many major earth-features, including the preponderance of land in the northern hemisphere, the relative positions of continents and oceans, the antipodeal relationships of land and sea, and the existence of the great south-pointing peninsulas such as South America and South Africa. Green suggested that a contracting earth with a relatively rigid crust would tend to assume that regular geometrical form, the tetrahedron, which has the largest area per unit volume. The continental blocks, which are elevations on the body of the earth, correspond with the corners of the tetrahedron, the oceans with its faces, although the deviation from a spheroid is, in fact, very small.[1] Processes involving the earth as a whole must be invoked to explain any world-wide system of genetically related structures, and thus Bucher[2] suggests that the pattern of the great mobile belts may be explained by crustal tension, which, he believes, alternated with compression according to variations in the volume of the subcrustal layers. On the other hand, several attempts have been made to recognize a world-wide pattern in features such as faults, fractures, and major relief forms, for example continental margins and submarine ridges. Such structures are termed in general *lineaments*. The existence of a pattern among lineaments over broad regions is readily seen—for example in the great rift valley systems (Fig. 42) or on a grander scale in the parallelism between the coasts of West Africa and eastern America, as also between these coasts and the Mid-Atlantic Ridge. From a statistical analysis of lineaments, Vening Meinesz postulated the existence of a global pattern forming a network in which two conjugate shear directions in the NE.–SW. and NW.–SE. quadrants are concerned.[3] He suggests a causal connexion with displacement of the poles, and on this premise deduces a hypothetical shear pattern,

[1] The theory is discussed by Bucher, W. H., *The Deformation of the Earth's Crust*: Princeton, 1933, pp. 464–8, and by Steers, J. A., *The Unstable Earth*: London, 1932.

[2] 'The Pattern of the Earth's Mobile Belts': *Journ. Geol.*, Vol. 32, 1924, pp. 265–90; also *op. cit.*, 1933, Chap. IV.

[3] 'Shear Patterns in the Earth's Crust': *Trans. Amer. Geophys. Union*, Vol. 28, 1947, pp. 1–61.

controlled by certain demonstrable lineaments, to which others should conform.

It is true that on the continents, major lineaments form a network in which two apparently conjugate trends are dominant, although others may also be present. This has been recognized in Europe,[1] Africa,[2] and Australia,[3] and is applicable also to other large areas.[4] But in the continents, and also in the oceans, meridional and east–west structure lines are important (as for instance in the Laurentian Shield, Central Australia, and the Urals) and these, as well as other important directions, are inadequately accounted for in Vening Meinesz' network.[5] Considerations that affect the establishment of any such world-pattern are inhomogeneity of the crust, including possible differences between large areas such as the Pacific Basin and the various continents and oceans, the existence of lineaments of different origin, some being tensional, others compressional; differences in age among lineaments and the recognition of trends due to secondary forces and effects. Although the structural geologist working in a small region

[1] Karpinsky, in a series of papers, was among the first to demonstrate the importance of conjugate lineaments. See 'Ocherki Geologicheskago Proshlago Evropeiskoi Rossyi': *Acad. Sci. U.S.S.R.*, Moscow, 1947 (Scientific Popular Series). Schwinner, R., 'Die Konsequenz in der tektonischen Entwicklung': *16th Internat. Geol. Congr.*, U.S.A., Rept. Vol. 2, 1936, pp. 983–92—demonstrated such lineaments in western Europe. See also Cloos, H., 'Grundschollen und Erdnähte': *Geol. Rundsch.*, Vol. 35, 1948, pp. 133–54; 'The Ancient European Basement Blocks': *Trans. Amer. Geophys. Union*, Vol. 29, 1948, pp. 99–103. Cizancourt, H. de, 'Quelques problèmes de tectonique géometrique': *Rev. Inst. Franc. du Pètrole*, Vol. 2, 1947, *passim*, arrives at an *orodynamic network* representing the distribution of the principal stress axes rather than of definable fractures or folds.

[2] Krenkel, E., *Geologie Afrikas*: Berlin, Vol. 1, 1925, Fig. 4. Cloos, H., 'Zur Grosstektonik Hochafrikas und seiner Umgebung': *Geol. Rundsch.*, Vol. 28, 1937, pp. 333–48.

[3] Hills, E. S., 'Some Aspects of the Tectonics of Australia': *Proc. Roy. Soc. N.S.W.*, Vol. 79, 1946, pp. 67–91. 'Tectonic Patterns in the Earth's Crust': *Pres. Addr. Sect. P, Aust. & N.Z. Assoc. Adv. Sci.*, Perth, 1947.

[4] For South America, see Von Estorff, F. E., 'Tectonic Framework of North Western South America': *Bull. Amer. Assoc. Petrol. Geol.*, Vol. 30, 1946, pp. 581–90.

[5] Umbgrove, J. H. F., *The Pulse of the Earth*: The Hague, 1947, pp. 304–8, gives a valuable discussion on the subject.

may be very much concerned with the latter, the recognition of broader structural controls is of fundamental importance, and may be of direct economic importance.[1]

Concerning the age of lineaments, although it is obvious that various major features such as geosynclines have shifted in position and varied in their mobility during geological time, almost all authors are agreed that a complex system of fractures affecting the earth was initiated at an early period, not long after the formation of a solid crust, and according to Cloos and others[2] the Pre-Cambrian basement is divided into parallel-sided blocks between which lie mobile *geosutures*. Much of the orogenic deformation of later time is localized in the geosutures, the trends of which thus influence younger structures. These latter are commonly parallel with the block-edges, but may also form obliquely, according to the direction of relative displacement of adjacent blocks (see pp. 61–3). The intricate trends of the European Alps, which combine rectilinear and strongly arcuate features, may be explained on this basis.

Resurgent Tectonics.—It is well known that faults of great antiquity may be rejuvenated under later stress, and in fact any large structural element in the crust, in which the rocks are markedly different from the surrounding masses, constitutes a feature that may influence structures of subsequent origin. Thus the application of idealized concepts such as the strain ellipsoid must be modified by inhomogeneities of the crust, which is very far from being an isotropic material. Ruedemann,[3] from a study of Pre-Cambrian trend lines (chiefly foliation and major fold trends) showed that the fundamental framework of continental structure is outlined by the

[1] See e.g. Billingsly, P. R., and A. Locke, 'Structure of Ore Districts in the Continental Framework': *A.I.M.E.* Spec. Pub. 51, 1939.

[2] Cloos, H., 'Zur Grosstektonik Hochafrikas und seiner Umgebung': *Geol. Rundsch.*, Vol. 28, 1937, pp. 333–48; 'Grundschollen und Erdnähte': *ibid.*, Vol. 35, 1948, pp. 133–54. Schwinner, R., 'Die Konsequenz in der tektonischen Entwicklung': *16th Internat. Geol. Congr.*, U.S.A., Rept. Vol. 2, 1936, pp. 983–92.

[3] 'The Existence and Configuration of Pre-Cambrian Continents': *N.Y. State Mus.*, Bull. 239–40, 1920–21, pp. 67–152.

trends of foliation, folds and shear zones, which have, presumably, influenced the younger structural lines (Fig. 27). This relationship receives a partial explanation on the hypothesis of basement blocks, since trends in the weak zones bordering these will for the most part parallel the blocks, and be repeated at various epochs including the later Pre-Cambrian. However,

FIG. 27.—PRE-CAMBRIAN TREND LINES IN AFRICA

(Compiled from various sources)

The chief trends of pre-Cambrian folding and foliation, with some important younger fold trends (dotted) are shown.

even Archaean trends are, with notable exceptions such as the oblique relationship of the Grenville belt to the Laurentian in Canada, reflected in later structures, as is well seen in India, Africa, and Australia.[1] It is notable, however, that where lineaments of different ages are not parallel, they are often at

[1] Hills, E. S., 'Tectonic Patterns in the Earth's Crust': *Pres. Addr. Sect. P., Aust. & N.Z. Assoc. Adv. Sci.*, Perth, 1947.

right angles or nearly so, oblique relationships being much rarer. This, and the sharp deflections in trends of all ages, are readily explicable if the basement blocks are conceived to be surrounded by partial frameworks of folds which lead to sharply divergent trends of the same or of different ages. While such relationships are reasonably based for many continental regions, other features, especially of the oceans, the great mobile belts and the Pacific margins, have geometrical properties directly related to surfaces that, it is now known, affect the subcrustal layers at least to a depth of 700 Km.

The observation of Sollas[1] that island and mountain arcs closely approximate to small circles of the globe was interpreted by Lake[2] to imply that they represent the intersection of thrust planes with the earth's surface. His views are generally substantiated by later geological and geophysical evidence,[3] seismological data on the depth of origin of earthquakes indicating that the foci lie in zones, which, as is especially clear in the western Pacific, dip beneath the continental margin and, at least in the crust if not in the substratum, are of the nature of thrust faults.[4] The centres of the small circles lie very close to two great circles, one circum-Pacific, the other corresponding with the Alpine-Himalayan mobile belt, and it seems that the smaller features may well be related to essentially vertical earth fractures on these great circles, the geometry of which has long been recognized. Boutakoff[5] demonstrates too that other lineaments, especially the volcanic lines of the North Pacific, and fracture belts such as the Red Sea, are arcs of great circles.

[1] Sollas, W. J., 'The Figure of the Earth': *Quart. Journ. Geol. Soc. London*, Vol. 59, 1903, pp. 180–8.

[2] Lake, P., 'Island Arcs and Mountain Building': *Geogr. Journ.*, Vol. 78, 1931, pp. 149–60.

[3] See e.g. Wilson, J. T., 'Some Major Structures of the Canadian Shield': *Canadian Min. & Met. Bull.*, Vol. 52, 1949, pp. 231–42.

[4] On the tectonic significance of the seismological data see Gutenberg, B., and C. F. Richter, *The Seismicity of the Earth and Associated Phenomena*: Princeton, 1949.

[5] 'The Great Circle Stress Pattern of the Earth, etc.': *18th Internat. Geol. Congr.*, London, 1948, Abstracts, p. 82.

In the extant pattern of the earth we may recognize firstly many features of great antiquity, especially in the continents. Younger structures are determined in part by the influence of these old trends, and in part by new tectonic influences. Many

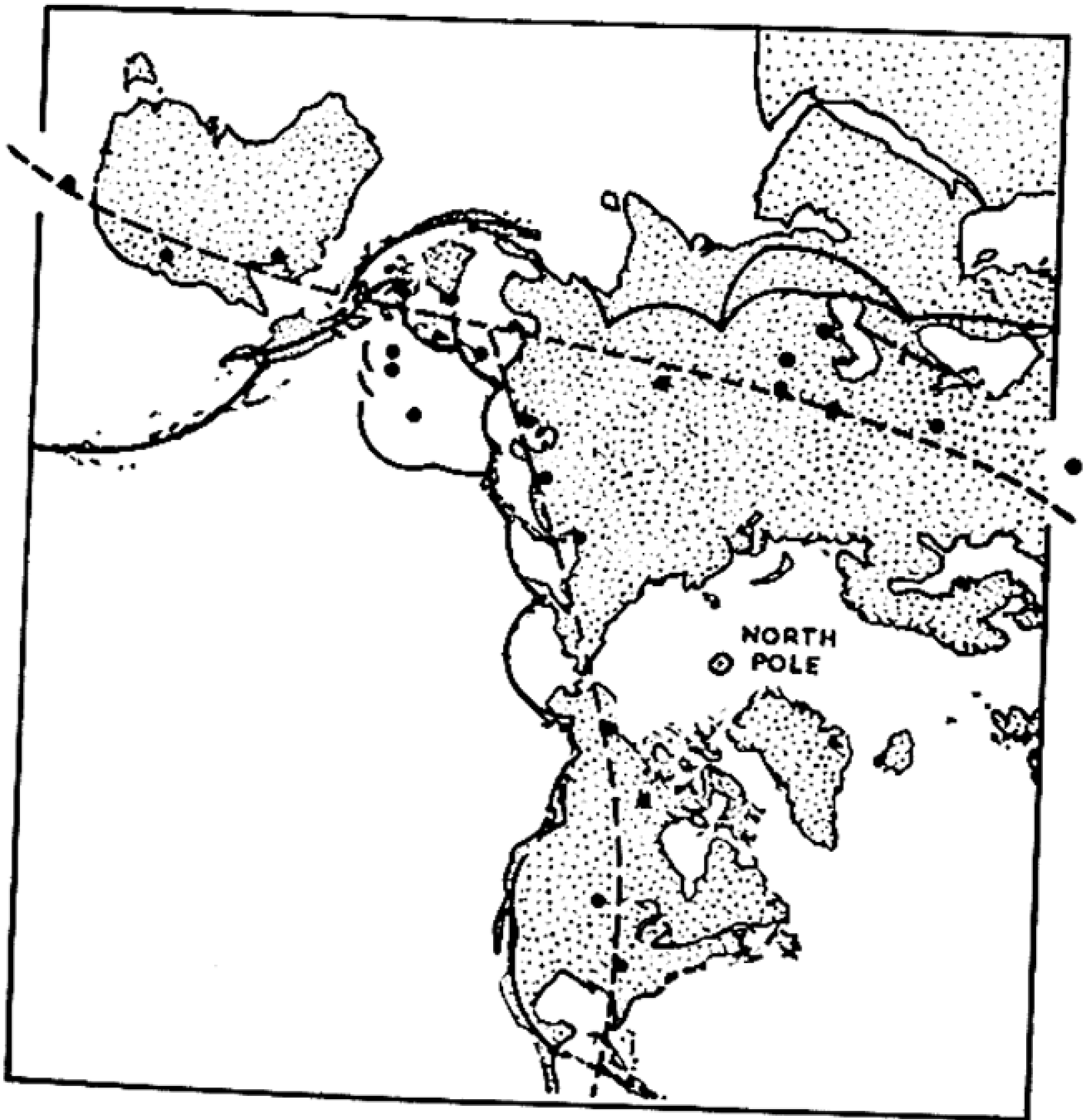


FIG. 28.—PATTERN OF FOLDED BELTS

(After J. T. Wilson)

Map on an oblique Mercator projection (Pole 35½° N. 2° E.) showing post-Triassic mountain and island arcs as small circles, their poles or centres, and two great circles through their poles.

major features, such as the young mobile belts and fracture lines in the oceans and continents, approximate to great circular arcs, and others to small circles. Continental margins have at all times been important tectonic features with which major crustal flexures and fractures, and geosynclinal troughs, have been linked.

3. *CHIEF TYPES OF CRUSTAL ARCHITECTURE*

The structures developed in a rock mass are determined by its physical properties and the system of forces that acts upon it. Owing to the general similarity in the nature of the rocks and of the forces that act upon them, in regions whose tectonic setting is broadly similar, comparable structures are developed in such regions in different parts of the world. For general descriptive purposes, and subordinating differences in detail, the chief types of crustal architecture may be classified as follows:

1. *Nappes* or *Decken.*
2. Normal folds—with or without thrust faults.
3. Plains type of folds.
4. Fault-folds (*Bruchfalten*).
5. Fault blocks.

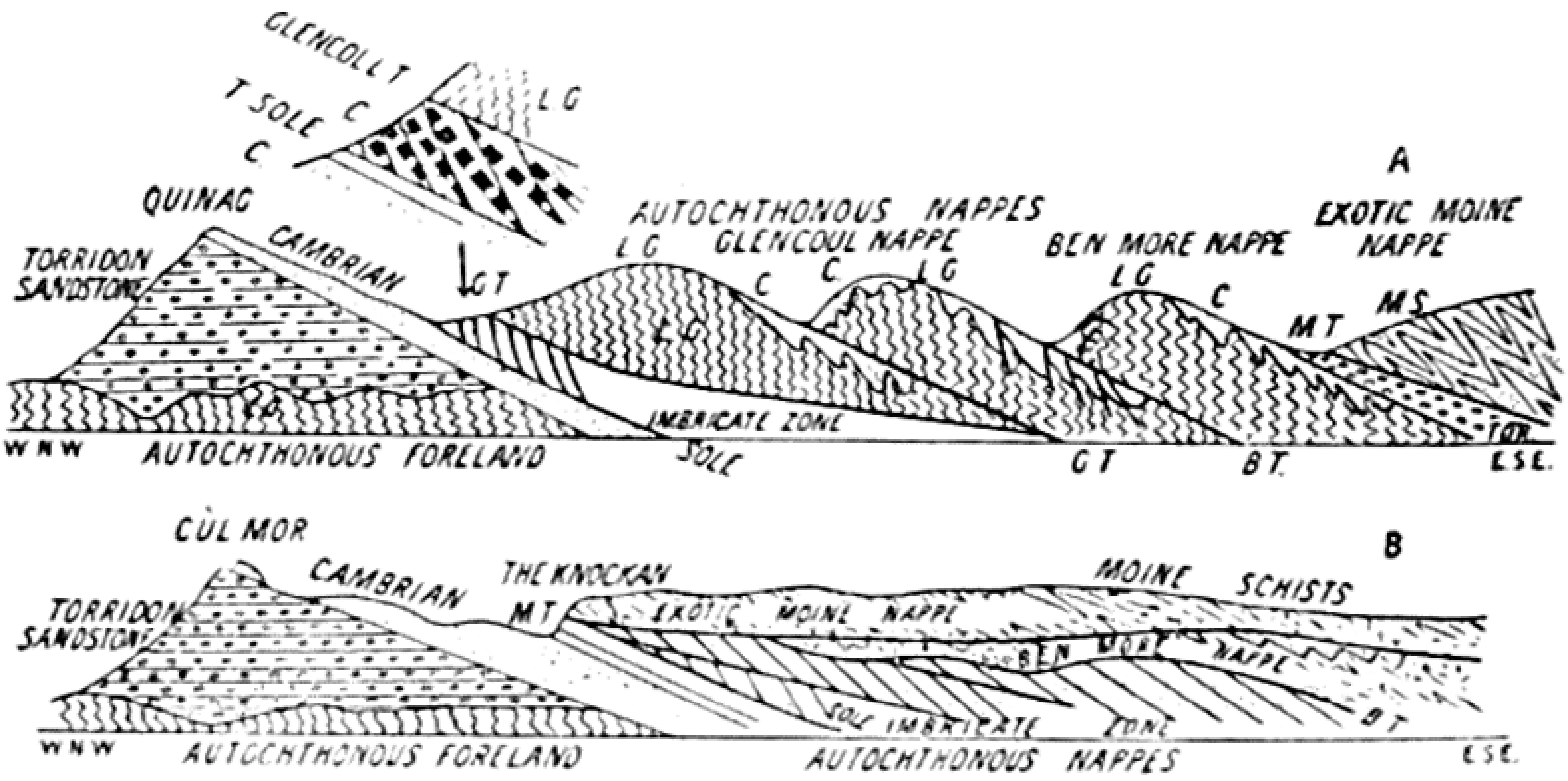


FIG. 29.—SKETCH SECTIONS ACROSS THE NORTH-WEST HIGHLANDS OF SCOTLAND, SHOWING THRUST NAPPES AND IMBRICATE STRUCTURE

(After Wills, *Physiographical Evolution of Britain*)

M.S., Moine Schists; C., Cambrian; TOR., Torridonian; L.G., Lewisian Gneiss; G.T., Glencoul Thrust; B.T., Ben More Thrust; M.T., Moine Thrust. Inset, details of the imbricate structure.

Nappe Structures.—A *nappe* (Fr.) or *Decke*[1] (Ger.) is a sheet of rocks, of large dimensions (of the order of miles), that has

[1] It should be noted that the terms *nappe* and *Decke* are used in French and German respectively for any covering sheet of rock, such as a layer of gravels or a basalt flow. In English, however, they are used only in a tectonic sense.

moved forward for a considerable distance (again of the order of miles) over the formations beneath and in front of it, finally covering them as a cloth covers a table. A nappe may be either the hanging wall (see p. 114) of a great low-angle overthrust (*thrust nappe, Überschiebungsdecke*; see Fig. 29), or a recumbent fold (*fold nappe, Überfaltungsdecke*; see Fig. 30), of which the

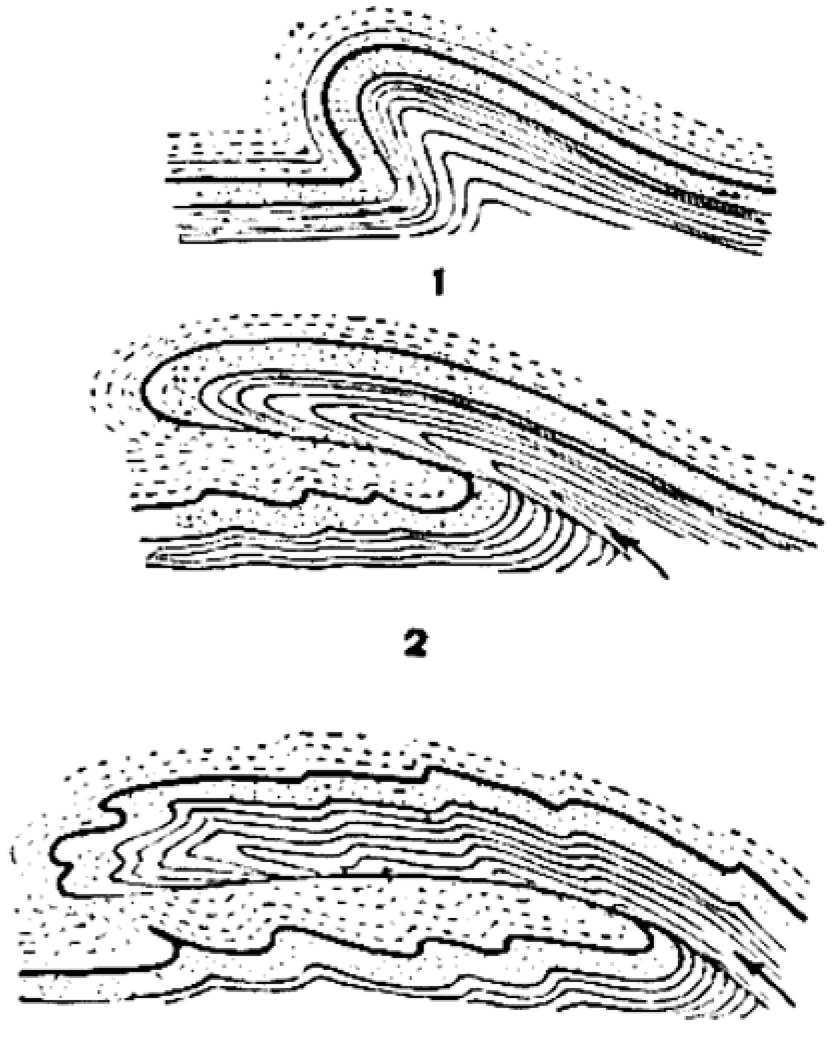


FIG. 30.—DEVELOPMENT OF A FOLD NAPPE BY THE SHEARING THROUGH OF THE MIDDLE LIMB OF A RECUMBENT FOLD

(After Heim, from Heritsch, *The Nappe Theory in the Alps*)

In Scottish Highland tectonics, a fault developing in the reversed limb of a fold is termed a *thrust*, and one replacing a normal limb a *lag* (see Bailey, E. B., *Quart. Journ. Geol. Soc.*, Vol. 90, 1934, p. 467; also *Mem. Geol. Surv. Scotland*, Explanation of Sheet 53, 1916, pp. 5–28).

reversed middle limb has been completely sheared out as a result of the great horizontal translation. Classical regions of nappe structure are the Highlands of Scotland (Fig. 29) and the European Alps (Fig. 31). Other well-known areas are the Rocky Mountains and the Himalayas.[1]

[1] For further information, see pp. 117–20, and the following works: EUROPEAN ALPS—Heritsch, F., *The Nappe Theory in the Alps* (translated by P. G. H. Boswell): London, 1929. Collet, L. W., *The Structure of the Alps*: London, 1st

In the Western Alps nappe structures have been deciphered in intimate detail, and a succession of tectonic zones differing in the type and intensity of deformation across the folded belt is clearly recognizable (see Fig. 31). Of the nappes themselves, the Pennine group originates in a *zone of roots*, where all rock structures are vertical or nearly so, and deformation is very

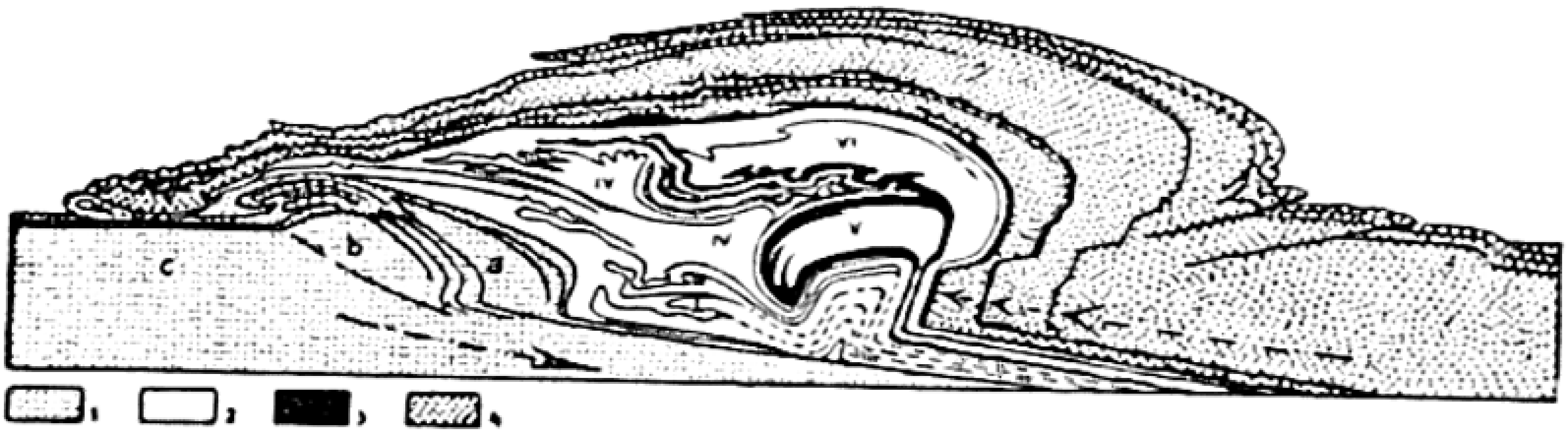


FIG. 31.—GENERALIZED SECTION SHOWING THE SUPPOSED STRUCTURE OF THE ALPS

(After Argand, from Collet, *The Structure of the Alps*)

1. *The Foreland or Eurasia.*—a, crystalline wedges; b, swelling of the crystalline basement; c, undeformed crystalline basement.
2. The Pennine nappes, which represent the western Alps.
3. Basic rocks.
4. *The Hinterland or Africa*, out of which the eastern Alps have been carved.

strong. The rocks of these nappes have travelled far from their original position, and are therefore termed *exotic*. North of the axial belt of Pennine nappes, a zone of Hercynian massifs, sheared into *crystalline wedges*, affords the roots of nappes in the

edn., 1927; 2nd edn., 1936. Bailey, E. B., *Tectonic Essays mainly Alpine*: Oxford, 1935. SCOTTISH HIGHLANDS—Peach, B. N., and J. Horne and others, 'The Geological Structure of the North-West Highlands of Scotland': *Mem. Geol. Surv. Great Britain*, 1907. Peach, B. N., and J. Horne, *Chapters on the Geology of Scotland*: London, 1930. Bailey, E. B., 'The Structure of the South-West Highlands of Scotland': *Quart. Journ. Geol. Soc.*, Vol. 78, 1922, pp. 82–131; 'The Glencoul Nappe and the Assynt Culmination': *Geol. Mag.*, Vol. 72, 1935, pp. 151–65; 'West Highland Tectonics: Loch Leven to Glenroy': *Quart. Journ. Geol. Soc.*, Vol. 90, 1934, pp. 462–525. Bailey, E. B., and H. B. Maufe, 'The Geology of Ben Nevis and Glen Coe': *Mem. Geol. Surv. Scotland*, No. 53, 1916. ROCKY MOUNTAINS—Flint, R. F., 'A Brief Review of Rocky Mountain Structure': *Journ. Geol.*, Vol. 32, 1924, pp. 410–30. Keith, A., 'Structural Symmetry in North America': *Bull. Geol. Soc. Amer.*, Vol. 39, 1928, pp. 321–86.

High Calcareous Alps that have not travelled far and are termed *parantochthonous*.

Further north in the Juras, and on the south in the Dinarides, thrusting and folding affect rocks that remain virtually in place as *autochthonous* features, although in the Pre-Alps exotic nappes override the autochthonous Mesozoic sediments.

FIG. 32.—SHOWING THE INDIVIDUAL FOLDS (SOMEWHAT GENERALIZED) IN THE FOLDED ZONE OF THE APPALACHIANS
(After Willis, 1893)

Normal Folding.—Associations of persistent, long and relatively narrow folds may be termed *normal fold structures*, in contrast with the recumbent and highly complex folds of the *Decken*.[1]

Typical examples of normal folding are afforded by the

[1] Seeing that the term *normal fold* is no longer in general use as a synonym for *symmetrical fold*, it might be used with advantage in the sense suggested above, without leading to confusion.

Appalachians and the Juras.[1] The Appalachians represent the deposits of a geosyncline, in which the strata have been thrown into long, narrow, parallel folds (see Fig. 32), often overturned and overthrust. The dominant type of fold is of the order of tens of miles long and single miles wide, 'sweeping in gentle curves with the trend of the belt' (Willis).

The Juras lie on the *foreland*[2] of the Alpine geosyncline. Their crystalline basement is not folded, but the overlying Mesozoic and Tertiary beds have been sheared over it and folded independently. This is the *décollement* of the folded Jura, which was facilitated by the presence of the plastic Anhydrite formation in the Middle Muschelkalk at the base of the sedimentary cover. Thrust faults occur, and the folds are usually subrectangular in cross-section, widely spaced anticlines being separated by flat-lying beds in the relatively undisturbed synclinal troughs (Fig. 33). This type of fold is termed a *Kofferfalt* (*box fold*).

As will be shown later (pp. 124–9), thrust faults develop by a variety of mechanisms in folded strata, so that regions of normal folding are often regions of thrust faulting. The faulting may take equal rank with the folding in structural significance, as in the Clowgill Burn section in the Ordovician rocks of Lanarkshire[3] (see Fig. 34). In less intensely folded beds, or with rocks of different physical properties, even if strongly folded, thrust faults may be absent.

Structure of Orogens.—Deformation connected with the Alpine orogeny in Europe exhibits well-defined tectonic zones parallel to the general trend of the orogenic belt. Such zoning

[1] Keith, A., 'Outlines of Appalachian Structure': *Bull. Geol. Soc. Amer.*, Vol. 34, 1923, pp. 309–80; 'Structural Symmetry in North America': *ibid.*, Vol. 39, 1928, pp. 321–86. Willis, B., 'Mechanics of Appalachian Structure': *13th Ann. Rept. U.S. Geol. Surv.*, Pt. 2, 1893, pp. 217–81. For the Juras see Collet, L., *The Structure of the Alps*: London, 1927, pp. 127–46, with full bibliography.

[2] The *foreland* is the resistant block towards which the geosynclinal sediments move when compressed: the *hinterland* is the actively moving block which forces them towards the foreland. If both blocks move equally this distinction breaks down.

[3] Peach, B. N., and J. Horne, 'The Silurian Rocks of Britain': *Mem. Geol. Surv.*, Vol. 1, 1899, p. 282.

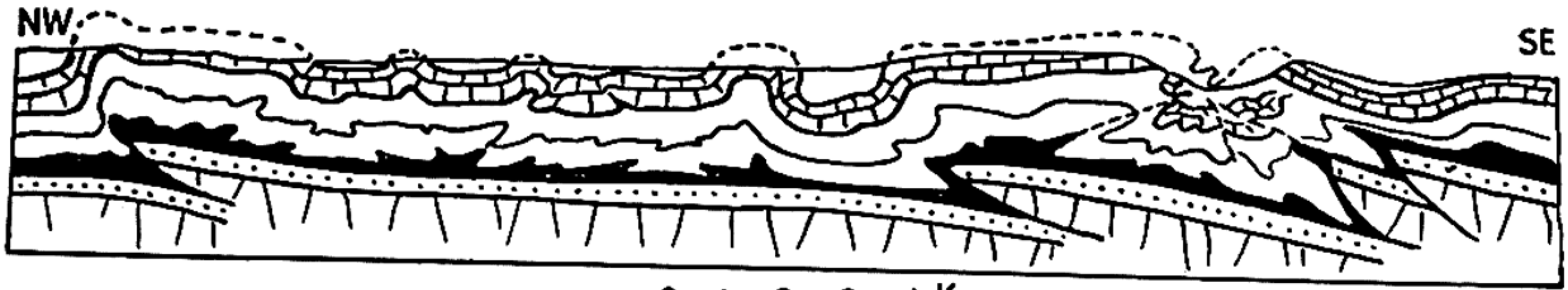


FIG. 33.—SECTION THROUGH THE JURAS

(After Aubert)

Shows the *décollement* of the sediments lying above the Anhydrite (solid black), and thrust faulting in the basement producing folds in the sedimentary cover.

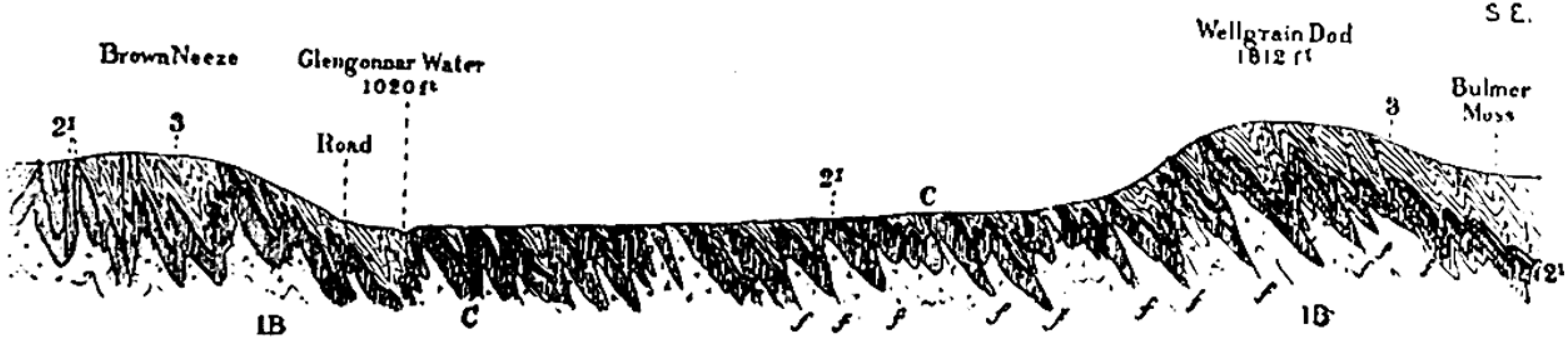


FIG. 34.—SECTION EXPOSED IN THE CLOWGILL BURN, GLENGONNAR WATER, LANARKSHIRE, SHOWING FOLDING AND THRUST FAULTING

(After Peach and Horne)

1B, Arenig volcanic rocks; C, Radiolarian chert; 2, Glenkiln shales; 3, Caradocian; *f*, fault. Length of section 1¼ miles.

(Reproduced from *Mem. Geol. Surv.*, 'Silurian Rocks of Britain, Vol. I, Scotland', 1892, by permission of the Controller of H.M. Stationery Office)

is a feature of all orogens, and according to Kober a symmetrical arrangement is normal, with bordering ranges on either side of a central, relatively undisturbed, region of *betwixt-mountains*, termed the *median mass*[1] (Fig. 35). In Peach's view the Scottish Highlands also show bilateral symmetry,[2] but the majority of deformed belts are asymmetrical, as in the western Alps (Fig. 31). Both types of structure are generally thought to have originated from horizontal compression of the geosynclinal sediments between the neighbouring rigid blocks, by a vice-like mechanism. This concept derives from the

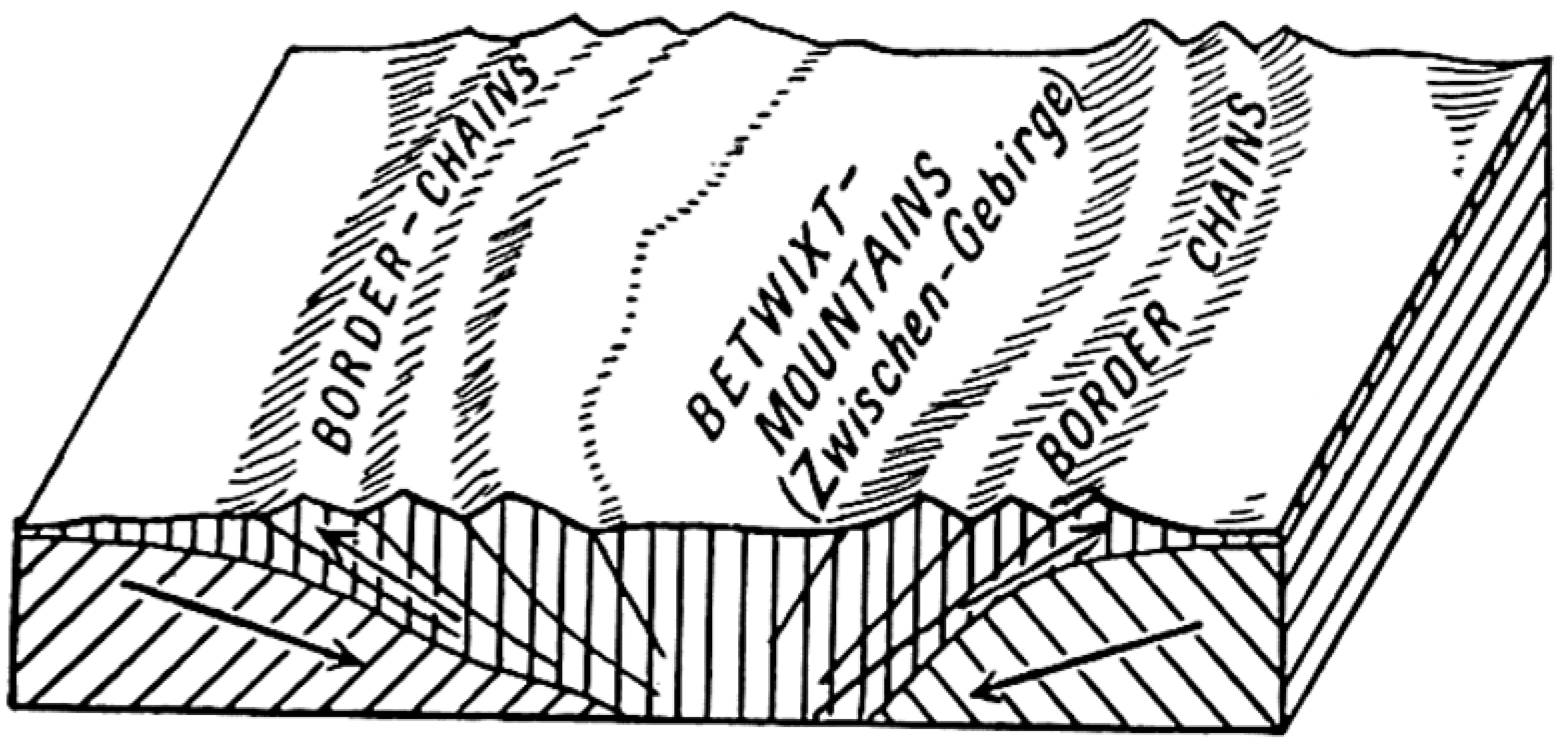


FIG. 35.—BLOCK DIAGRAM OF PORTION OF A TYPICAL OROGEN ACCORDING TO KOBER

(After Wills, *Physiographical Evolution of Britain*)

classic experiments of Hall on folding,[3] but it is also suggested that folding may result from horizontal shearing movements between the resistant blocks bordering the geosyncline. Experimental studies on the development of fold structures have shown that folding can result from such movements,[4] but the

[1] Kober, L., *Der Bau der Erde*: Berlin, 1928, p. 173.

[2] Peach, B. N., and J. Horne, *Chapters on the Geology of Scotland*: Oxford, 1930, Fig. 27.

[3] Hall, J., 'On the Vertical Position and the Convolutions of certain strata, and their relation with Granite': *Trans. Roy. Soc. Edinburgh*, Vol. 7, 1815, p. 79.

[4] Mead, W. J., 'Notes on the Mechanics of Geologic Structures': *Journ. Geol.*, Vol. 28, 1920, pp. 505–23. Leith C. K., *Structural Geology*: New York (revised edn.), 1923, p. 192.

arrangement of the folds within the deformed zone is characteristically different from that in a zone of soft rocks deformed by direct compression (see Fig. 36). In folding by direct compression, the direction of greatest relief is upwards, the least strain axis is in the direction of the active forces, and the mean strain axis is horizontal and at right angles to this direction. The fold axes are parallel to each other, and at right angles to the deforming forces. They are relatively long, and in geosynclines

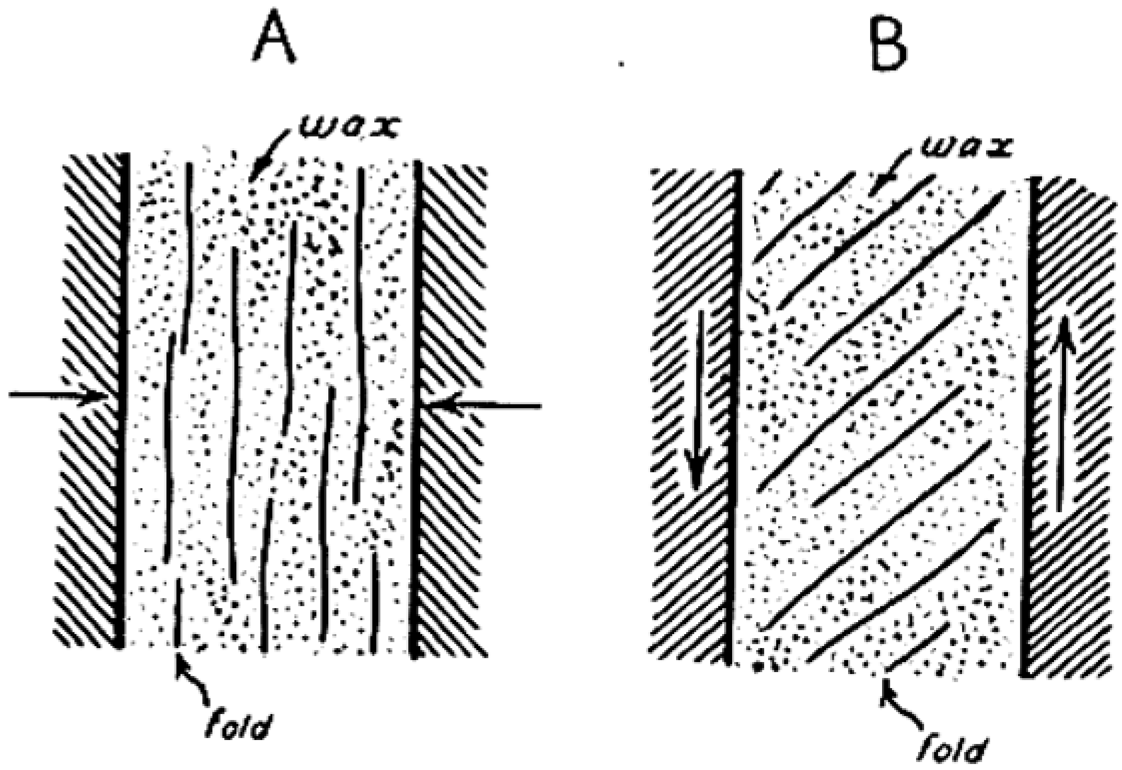


FIG. 36.—FORMATION OF FOLDS IN PARAFFIN WAX

(After Mead, 1920)

A. By direct compression: the axial lines of the folds are parallel to the edges of the deformed sheet of wax, and at right angles to the direction of compression.

B. By horizontal shearing stress: the axial lines are arranged *en echelon*, and lie at 45° to the borders of the wax sheet.

would be parallel to the margins. On the other hand, experimental studies have shown that in material subjected to shearing stress, folds are produced which are arranged *en echelon* within the deformed mass, with the fold axes at 45° to the direction of the shearing couple, and at right angles to the direction of shortening[1] (see Fig. 36). Thus, if the resistant

[1] Hubbert, M. K., 'The Direction of the Stresses producing given Geologic Strains': *Journ. Geol.*, Vol. 36, 1928, pp. 75–84. Chamberlin, R. T., 'The Strain Ellipsoid and Appalachian Structures': *ibid.*, pp. 85–90. Mead, W. J., 'Notes on the Mechanics of Geologic Structures': *Journ. Geol.*, Vol. 28, 1920, pp. 505–23.

blocks on either side of a geosyncline move relatively to each other in a direction parallel with the axis of the geosyncline, a similar arrangement of folds *en echelon* would be expected to develop. This is not found in the Appalachians, where the folds are long and parallel to the borders of the geosyncline, so that, in all probability, the folding there was due to direct compression.

The importance of horizontal shearing movements has, however, been urged by those who have studied the young mountain chains of the western Pacific regions, where the arrangement of the mountain arcs is readily explained by combinations of direct compression, horizontal shear, and torsion about a vertical axis.[1] The various patterns of folds observed in eastern Asia have been reproduced by Lee in experiments with softened tracing paper manipulated in various ways on a polished board.[2] The chief types of pattern he recognizes are shown in Fig. 37.

It is realized, moreover, that a simple vice-like mechanism is inadequate to explain folding in a belt some scores or even hundreds of miles in width, since soft rocks could not transmit stress over such distances; and again, the compact basement on which the sediments rest cannot fold in the same style as these, so that some different response of the basement rocks to stress has to be postulated and accounted for. In this connexion, Argand[3] visualizes a general plastic flux in the basement, which yields by *plis du fond* and by low-angle thrusts, in response to which the cover deforms differently, by *plis de couverture*. The Juras afford a classic example of a sharp distinction between superficial and deep tectonics, where the basement is still rigid, and Aubert[4] has argued that the local zones of strong folding in the Juras originate above high-angle thrusts in the crystalline

[1] Tokuda, S., 'On the Echelon Structure of the Japanese Archipelagoes': *Jap. Journ. Geol. Geog.*, Vol. V, 1926–7, pp. 41–76.

[2] Lee, J. S., 'Some Characteristic Structural Types in Eastern Asia and their bearing on the Problem of Continental Movements': *Geol. Mag.*, Vol. 66, 1929, *passim*.

[3] Argand, E., 'La Tectonique de l'Asie': *Congr. Geol. Internat.*, Session 13, Belgium, 1922 (1924).

[4] Aubert, D., 'Le Jura': *Geol. Rundsch.*, Vol. 37, 1949, pp. 2–17.

basement (Fig. 33). The origination of folds in buried thrust blocks is recognized in the autochthonous nappes of the Alps

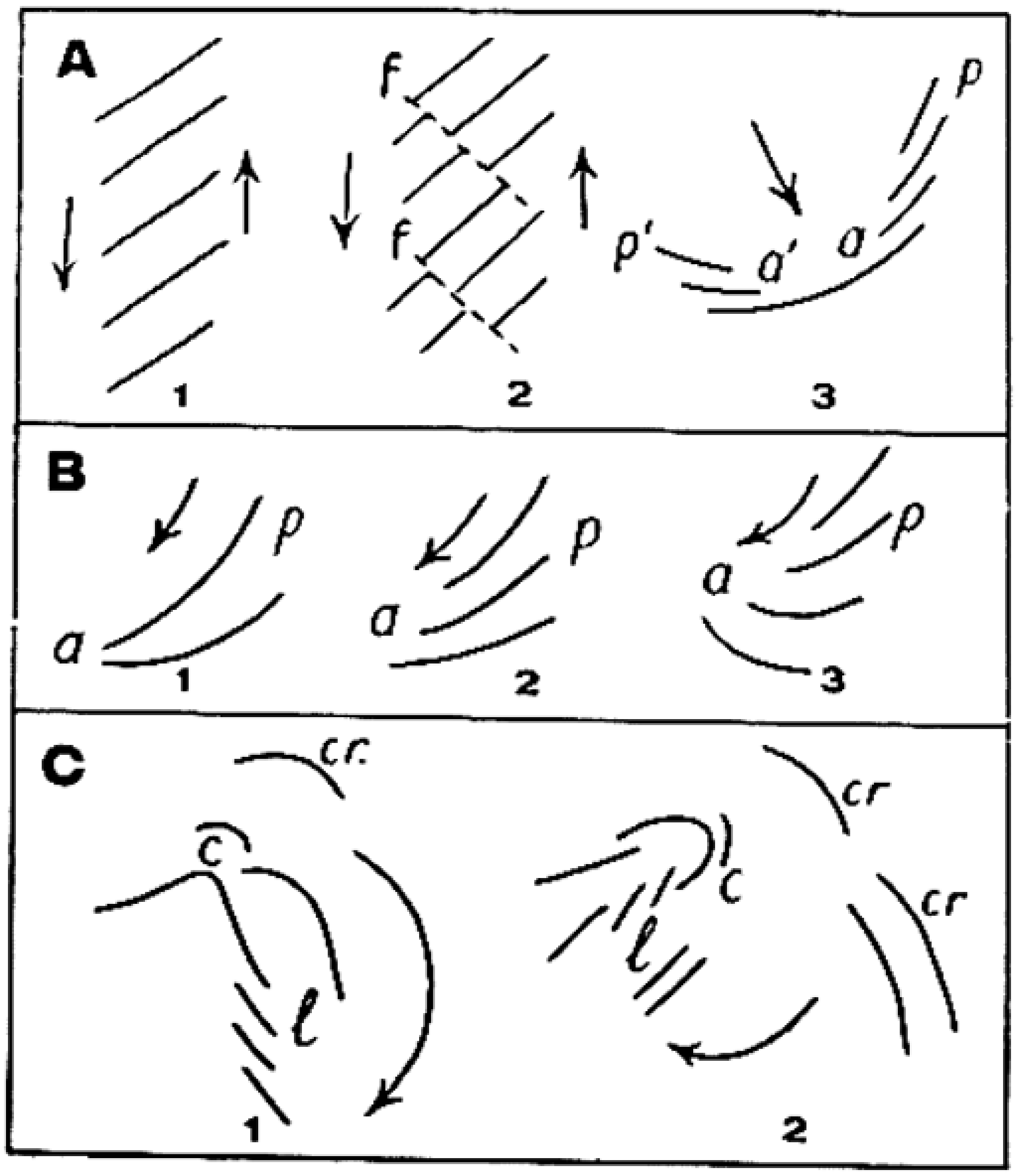


FIG. 37.—CLASSIFICATION OF STRUCTURAL TYPES ACCORDING TO THE ARRANGEMENT OF MAJOR FOLDS AND THEIR PROBABLE MODE OF ORIGIN. THE ARROWS INDICATE THE PRINCIPAL DIRECTIONS OF MOVEMENT

(After Lee, 1929)

A. Different forms of xi (Greek ξ) structure.
1, parallel xi type; 2, parallel xi type with tear faults f, f; 3, convergent xi type: a and a′, anterior ends, p and p′, posterior ends.

B. Different forms of nu (Greek ν) structure.
1, Simple nu; 2, ordinary type of nu; c, spiral-like type.

C. Different forms of eta (Greek η) structure.
1, elongate form; 2, broad form; c, cranium: Cr, crown: 1, limb.

and in New York,[1] and as a general concept has much to commend it. The connexion of folds with normal faults is also recognized (see pp. 67–9).

In Argand's synthesis of the western Alpine structure, the required shortening of the crystalline basement is achieved by

[1] Balk, R., 'Structural and Petrologic Studies in Dutchess County, New York. Pt. 1': *Bull. Geol. Soc. Amer.*, Vol. 47, 1936, pp. 685–774.

major overthrusting of the African block over the European block, so forming a great sledge (*traineau ecraseur*). The concept of over (or under) thrusting is inherent in many theories of orogenesis[1] and receives support from seismological evidence, especially for young mountain arcs marginal to continents.

Another theory of tectogenesis, based largely on the study of gravity anomalies in the East and the West Indies, accounts in an entirely different way for the shortening of the basement by the postulate of a great downward buckle of the granitic layer which becomes isoclinally folded to form a *tectogene* while the sedimentary cover is skimmed off it and develops complex structures independently.[2] Experimental studies by Griggs support the hypothesis, and also indicate that the downbuckle may be formed as a result of the drag of subcrustal convection currents on the base of the granitic layer (Fig. 38, A). A mechanism involving convection currents and the downward drag of the granite layer is also postulated by Holmes.[3]

In the theories outlined above, lateral displacements of the crust of the order of many miles are postulated, although of lesser amounts than are conceived in the hypothesis of continental drift. Other theories of diastrophism are markedly contrasted in that the movements of the crust involved are chiefly vertical and the development of orogenic structures in the sedimentary cover is achieved by secondary tectonics, chiefly by sliding under the pull of gravity.

Gravitational Tectonics.—The notion that the complex folds and faults of orogenic belts may have formed by the sliding of a sedimentary cover over an inclined basement under the action of gravity was first expressed in the nineteenth century, but did not gain favour until recent years, when Jeffreys and Bull in England[4] and Lugeon and Gagnebin on the

[1] See e.g. Lawson, A. C., 'Insular Arcs, Fore Deeps, and Geosynclinal Seas of the Asiatic Coast': *Bull. Geol. Soc. Amer.*, Vol. 43, 1932, pp. 353–81.

[2] Vening Meinesz, F. A., 'Gravity Expeditions at Sea': *Netherlands Geodetic Comm.*, Vol. II, 1934, Delft. Griggs, D., 'A Theory of Mountain-Building': *Amer. Journ. Sci.*, Vol. 237, 1939, pp. 611–50.

[3] *Principles of Physical Geology*: London, 1945.

[4] Jeffreys, H., *Earthquakes and Mountains*, London, 1935. Bull, A. J., 'The Compression of a Sheet': *Proc. Geol. Assoc.*, Vol. 54, 1943, pp. 185–90.

Continent[1] have propounded the theory for the Alps and the Juras, and the concept is also incorporated in general theories —the Oscillation Theory of Haarmann[2] and the Undation Theory of van Bemmelen.[3]

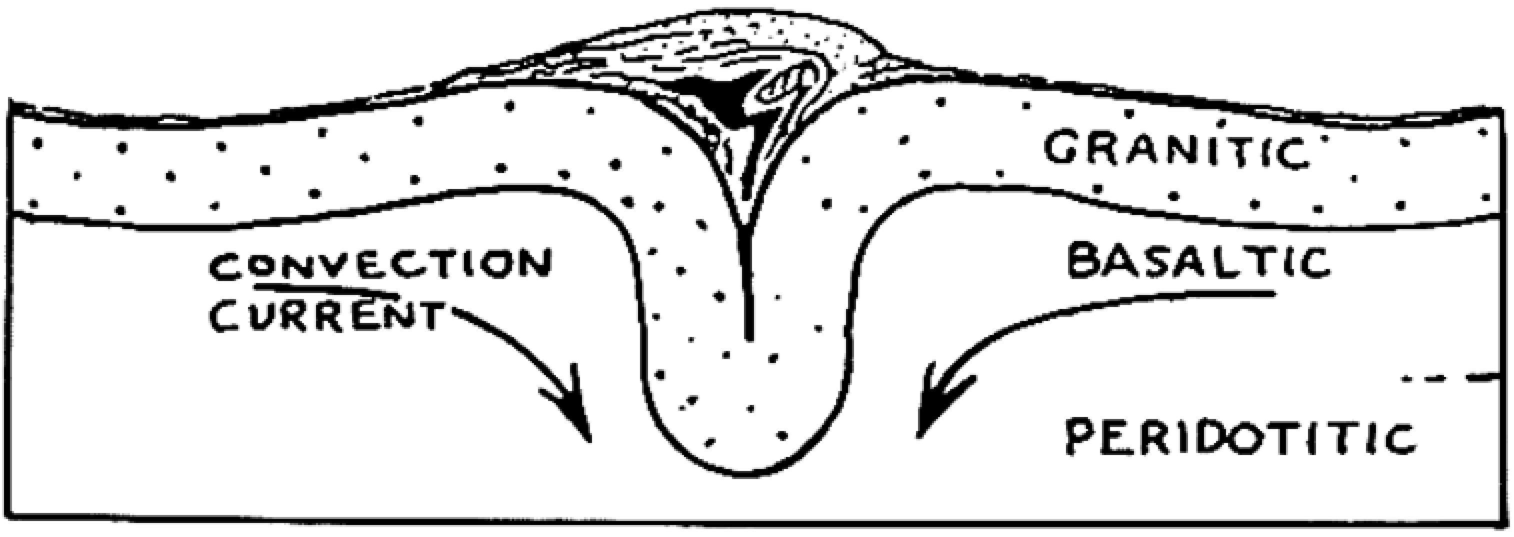


FIG. 38 (A).—CRUSTAL DOWNBUCKLE OR TECTOGENE

(After Hess and Griggs)

Shows the buckled granitic layer, the western Alps superimposed thereon to the same scale, and suggested convection currents.

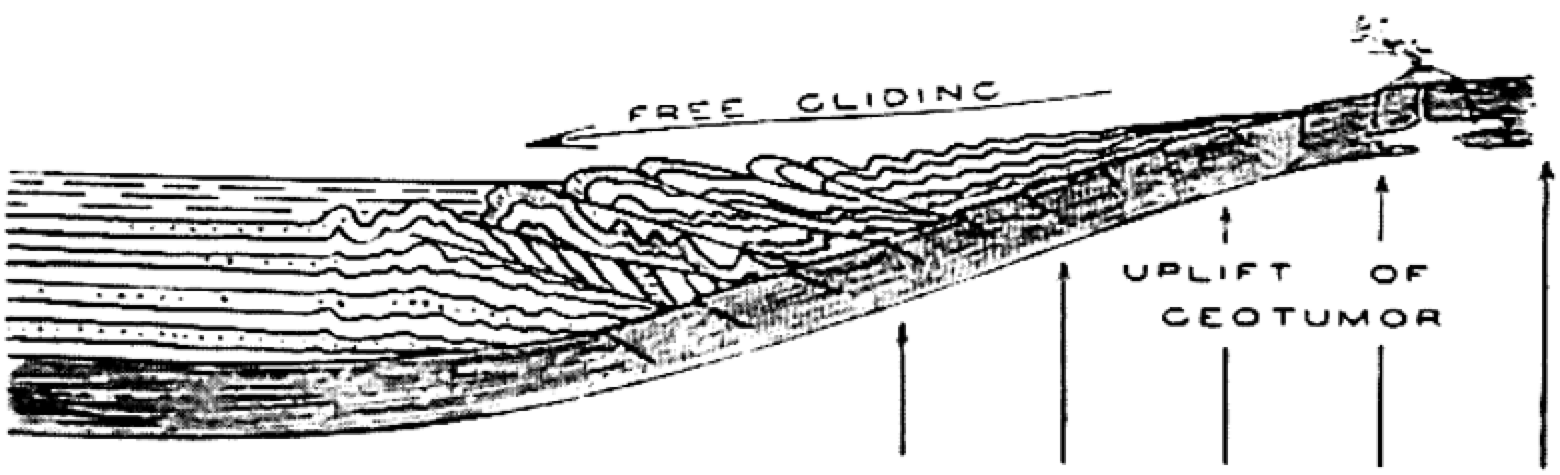


FIG. 38 (B).—DEVELOPMENT OF AN ASYMMETRICAL ARRANGEMENT OF STRUCTURES IN A MOBILE BELT BY FREE GLIDING

(After Haarmann)

In general, the above-mentioned authors postulate uplift and tilting of the basement rocks, together with their sedi-

[1] For reviews of the subject see Gignoux, M., 'La Tectonique d'Écoulement par Gravité et la Structure des Alpes': *Bull. Soc. Géol. France*, Ser. 5, Vol 18, 1948, pp. 739–61. Also Symposium in *Geol. en Mijnbouw*, Jhg. 12, N.S., No. 12, 1950; Gignoux, M., 'Meditations sur la Tectonique d'Écoulement par Gravité: *Trav. Lab. Géol. Univ. Grenoble*, Vol. 27, 1948, pp. 1–34; and Gagnebin, E., 'Quelques Problèmes de la Tectonique d'écoulement en Suisse Orientale': *Bull. Lab. Géol. Univ. Lausanne*, No. 80, 1945.

[2] Haarmann, H., *Die Oszillations-Theorie*: Stuttgart, 1930.

[3] van Bemmelen, R. W., *The Geology of Indonesia*: The Hague, 1949.

mentary cover, accompanied and followed by plastic flowage of the cover (*Freigleitung* of Haarmann) (Fig. 38, B). Although many structural features such as fold-nappes and the pattern of folds in plan are readily explainable on this hypothesis, thrust-faulting in the crystalline basement and the strongly developed vertical tectonics in regions such as the zone of roots in the Alps require compression. It is, however, the invocation of a force (gravity) that affects every particle of a rock mass and does not require the transmission of stress by the deforming rocks themselves that perhaps appeals most in the theory of gravitational tectonics, and as Kaisin[1] has pointed out for the Ardennes, a tectonic field of force (*champ tectonique*) of all-pervading nature must be postulated for many orogens.

An alternative view involving gravitational flowage is expressed by Fourmarier,[2] who points out that such flow may well have taken place in the geosynclinal stage rather than later. This view is supported by geosynclines in which the intensity of structural development (folds and faults) increases progressively in the older and deeper rocks, a feature that is classically shown in the Ruhr, and is referred to as Böttcher's Principle. Various explanations of the Ruhr folding have been proposed.[3]

In U.S.A., recent interpretations have demonstrated the importance of lateral translation of several miles, in some instances along flat lying bedding faults.[4] These ideas are incompatible with simple lateral compression.

[1] Kaisin, F., 'Poussées tangentielles ou 'Champs Tectoniques'?: *Bull. Soc. belge Géol.*, Vol. 53, 1944, pp. 228–63.

[2] Fourmarier, P., 'Efforts tangentiels et efforts verticaux dans la tectogenèse': *Bull. Soc. Géol. de Belg.*, Vol. 69, 1946, pp. 87–182.

[3] Böttcher, H., 'Stockwerkstektonik oder Faltungstiefenstufe im Gebiet der Ruhrmolasse': *Glückauf*, Vol. 80, 1944, pp. 9–14. Oberste-Brink, K., 'Sedimentation und Tektonik im Ruhrkohlenbezirk': *ibid.*, Vol. 69, 1933, p. 693. Kober, L., 'Zur Geotektonik der Ruhrmolasse': *ibid.*, Vol. 79, 1943, pp. 469–75.

[4] Rogers, J., 'Evolution of Thought on Structure of Middle and Southern Appalachians': *Bull. Amer. Assoc. Petrol. Geol.*, Vol. 33, 1949, pp. 1643–54. also 'Mechanics of Appalachian Folding': *ibid.*, Vol. 34, 1950, pp. 672–81. Longwell, C. R., 'The Mechanics of Orogeny': *Amer. Journ. Sci.*, Vol. 243A; 1945, pp. 417–47.

When the range of possible causes of tectogenesis is considered, it is clear, as recent work shows, that structural interpretation must be based upon the apparent tectonic setting and should not be in any way restricted by over-simplified notions concerning the mechanism of diastrophism.[1]

Plains Type of Folding.—In the Mid-Continent region of North America, fold structures occur which have the following characteristics.[2] (1) The net result of the folding is a local uplift without a corresponding depression, so that there are

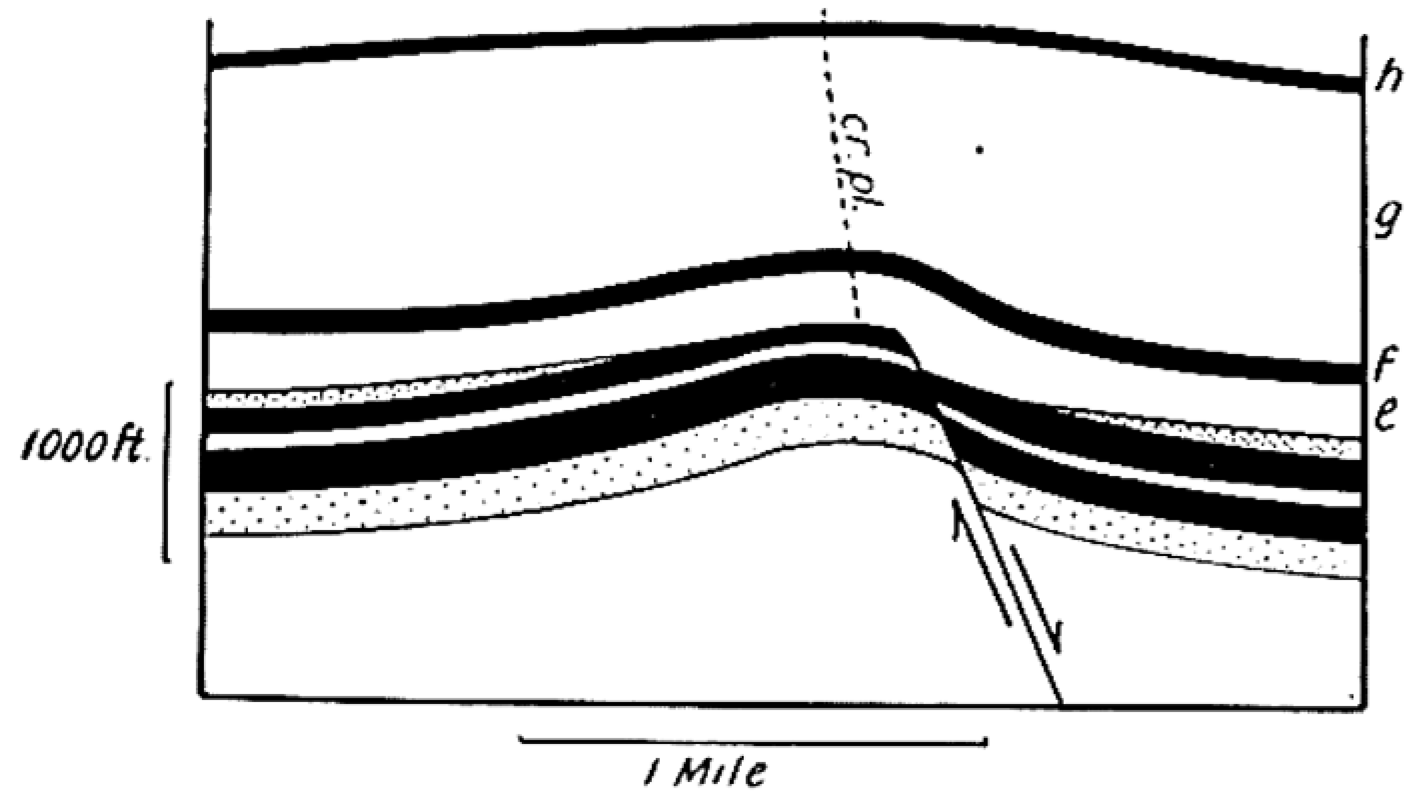


FIG. 39.—FOLDING OF THE PLAINS TYPE, DEVELOPED IN THE BEDS e, f, g, h, BY FAULTING IN THE BASEMENT UPON WHICH e RESTS UNCONFORMABLY

(After Clark, 1932)

Cr. pl., trace of crestal plane.

only anticlines and domes rising 'above' the regional dip; (2) the folds become more pronounced in depth, and there is a thinning of the strata above the crests of the folds; (3) the folds are usually asymmetrical, and the crestal plane (see p. 67) dips towards the steeper limb of the fold; (4) normal faulting is commonly associated with the folding (see Fig. 39).

[1] See e.g. Gray, K. Washington, 'A Tectonic Window in South-western Iran': *Quart. Journ. Geol. Soc.*, Vol. 105, 1950, pp. 189–224, in which the existence of a nappe that has been projected for 30 miles over a flat and undisturbed surface is postulated.

[2] Clark, S. K., 'The Mechanics of the Plains-Type Folds of the Mid-Continent Area': *Journ. Geol.*, Vol. 40, 1932, pp. 46–61.

The development of folds of this type is believed to be due in part to differential movements in the bedrock upon which the folded sediments lie, and not to lateral compression acting on the sediments themselves. The beds were deposited upon an uneven surface, and later stress has accentuated the differences in elevation of the various parts of this surface. The buried hills are, in some examples at least, upthrown fault blocks which on renewed uplift, after having been covered with sediments, exert an upward push which gives rise to the anticlines and domes, some of these having marginal normal faults. The geometry of the folds is determined by the initial form of the bedrock surface, the extent to which the buried hills are rejuvenated, and the compaction and flowage of the sediments during their deformation[1] (pp. 85–6).

Fault-folding (Bruchfaltung).—Stille and others[2] have shown that parts of middle and north Germany (Saxony) are characterized by a close association of folding with nearly vertical faults, which is described as *fault-folding*.[3] The crust is divided into a number of long, narrow, differentially elevated strips, the upthrown blocks representing anticlinal structures, the limbs of which are affected by faults striking parallel to the length of the folds (see Fig. 40). The sediments in the upthrown blocks show little or no folding, but in the downthrown blocks

[1] Clark, S. K., 'The Mechanics of the Plains-Type Folds of the Mid-Continent Area': *Journ. Geol.*, Vol. 40, 1932, pp. 46–61. Powers, S., 'Structural Geology of the Mid-Continent Region; a Field for Research': *Bull. Geol. Soc. Amer.*, Vol. 36, 1925, pp. 379–92; also 'Structural Geology of North-Eastern Oklahoma': *Journ. Geol.*, Vol. 39, 1931, pp. 117–32. Ferguson J. L. and J. Vernon, 'Relationship of Buried Hills to Petroleum Accumulation': *The Science of Petroleum*: Oxford, 1938, pp. 240–3.

[2] Stille, H., 'Mitteldeutsche Rahmenfaltung': *Jahrb. d. niedersächs. Ver. Hannover*, 1910, pp. 141–69; also, with others, in 'Göttinger Beiträge zur saxonischen Tektonik': *Abh. Preuss. Geol. Landesanstalt*, N.F., Vol. 95, 1923–5, Vol. 116, 1930, Vol. 128, 1931. 'Geotektonische Forschungen': Hft. 1, *Zur Germanotypen Tektonik I.* Edited by H. Stille and F. Lotze, Berlin, 1937.

[3] The term *fold-fault* has long been used for faults developed in close connexion with folds, especially in nappe tectonics (see Bailey, E. B., *Quart. Journ. Geol. Soc.*, Vol. 90, 1934, p. 467). The restriction of the use of the similar term *fault-fold* to *Bruchfalten* of the Saxonian type may therefore lead to confusion, but a better word has not yet been suggested.

the intensity of the folding increases, as does also the thickness of sediments. A few overthrusts are present, and there are many local unconformities.

The Coast Ranges of California afford another example of a region exhibiting this type of structure,[1] which Bucher interprets as being caused by alternating tension and compression in an area where thick deposits of sediments are being laid down. According to this view, the pattern of the long blocks bounded by normal faults is determined by tension. The folding of the sediments in the troughs, as well as the minor thrusting, result from later compression and from drag

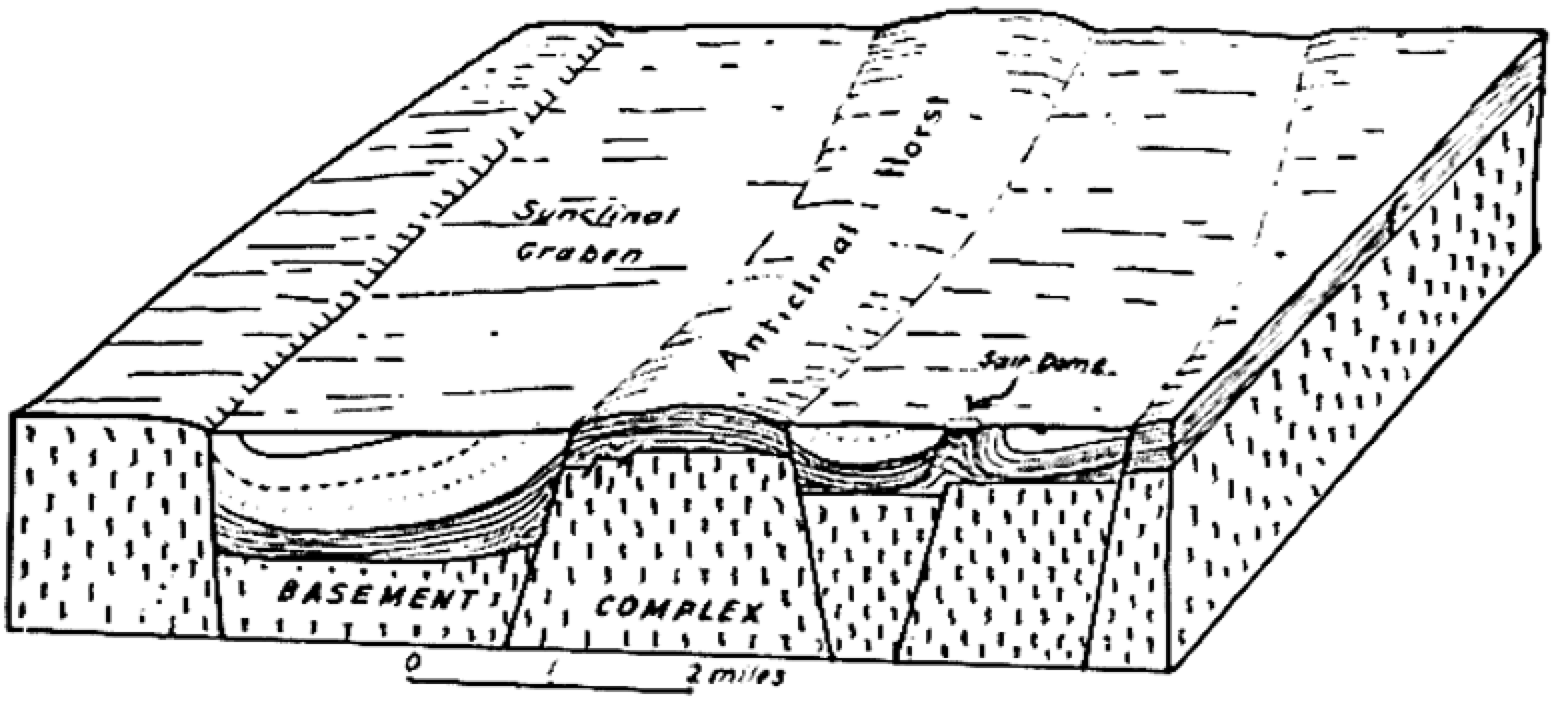


FIG. 40.—BLOCK DIAGRAM TO ILLUSTRATE BRUCHFALTEN

Salt deposits are shown at the base of the sedimentary cover lying on the basement complex.

along the faults. Schuh[2] came to much the same conclusion with regard to the Saxon *Bruchfalten*, which Stille considered to be caused by compression alone, despite the dominance of normal faulting. Structures comparable with *Bruchfalten* are probably much commoner than would appear from the literature, many examples being described in terms of faulting, with drag effects, rather than in terms of folding. There is a close similarity with folding of the Plains type.

[1] Bucher, W. H., *The Deformation of the Earth's Crust*: Princeton, 1933, pp. 303–25. Clark, B. L., 'Folding of the California Coast Range Type illustrated by a Series of Experiments': *Journ. Geol.*, Vol. 45, 1937, pp. 296–319.

[2] Schuh, F., 'Die Saxonische Gebirgsbildung,: *Kali*, Vol. 16, 1922, Hfte. 8, 9, 15, 16.

Block Faulting.—Regions which are divided by faults into a number of differentially elevated or depressed blocks are said to exhibit *block faulting*.[1] If such a region has not been subjected to prolonged erosion, the movements of the blocks will be directly reflected in the topography. Upstanding blocks, either plateaux or ridges, are termed *horsts*, and blocks which have received a marked tilt, *tilt blocks*. Normal faults (see pp. 115–17) that give rise to a series of blocks at successively lower or higher elevations are termed *step faults*, and long and relatively narrow tectonic valleys caused by faulting are called *fault troughs* or *grabens*.

If the major movements take place along only one set of sub-parallel faults, long and relatively narrow blocks result. A second set of faults, intersecting the others, may cause the boundaries of the troughs and horsts to assume a zig-zag pattern in plan, as in East Africa; or, the crust may be divided into a number of small blocks, constituting a *block mosaic*.[2]

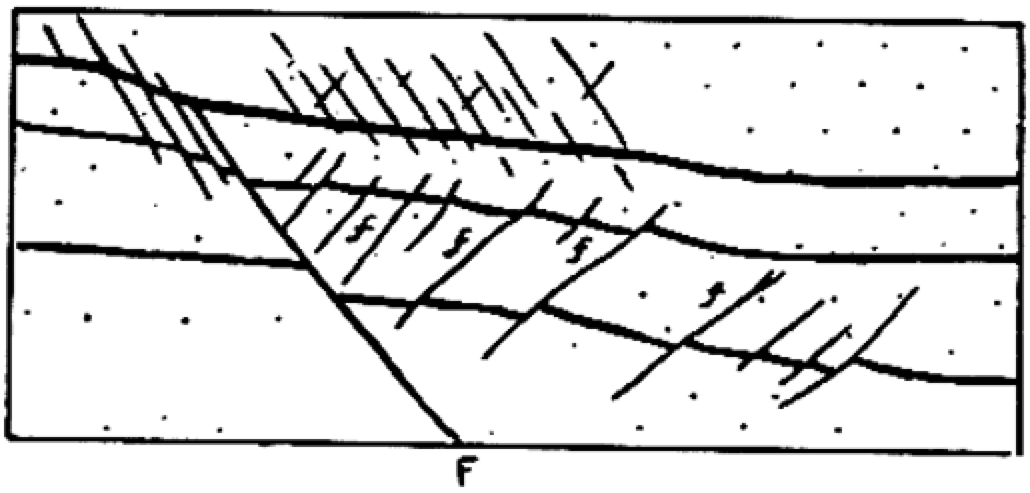
The Origin of Grabens.—If a faulted area has been subjected to prolonged erosion, geological mapping will generally reveal whether the faults are normal or reverse (see pp. 115–17), but where the faulting is physiographically youthful it is more difficult to determine this. In particular, the structure and origin of the large grabens of East Africa, the Dead Sea region, and the Rhine valley are still subjects of debate in spite of the efforts of numerous workers.

The Rhine graben is known in most detail, and concerning it there is least doubt as to the structure. The graben traverses a broadly arched region that comprises the Vosges and the Schwarzwald in the south and the Hardt and the Odenwald in the north. The boundary faults are normal faults, hading towards the graben, and the great majority of minor faults within the graben are also normal faults. H. Cloos has reproduced the structural features of this graben so faithfully in experiments with wet clay subjected to lateral tension (see

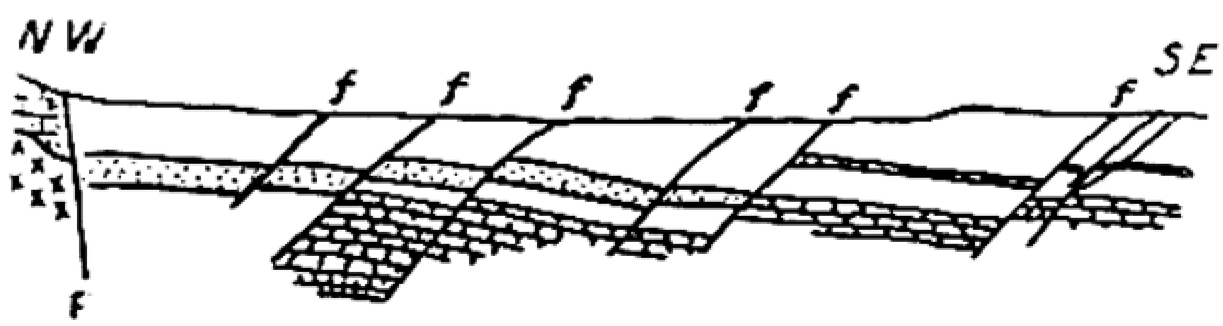
[1] The term is not usually applied if the faults are of the strike-slip variety (see p. 114).

[2] As in central Sweden; see Davis, W. M., *Die Erklärende Beschreibung der Landformen*: Berlin, 2nd edn., 1924, p. 169.

Fig. 41, and Plate IV), that doubt can hardly be entertained as to its origin. It was undoubtedly caused by circumferential tension in the crust at right angles to the graben, which may



A



B

FIG. 41.—STRUCTURE OF PORTION OF AN EXPERIMENTALLY PRODUCED GRABEN IN CLAY (A), FOR COMPARISON WITH PART OF THE WESTERN BORDER OF THE RHINE GRABEN (B)

(After Cloos; redrawn from Bucher, *The Deformation of the Earth's Crust*)

The shearing plane F in the clay is comparable with the boundary fault F of the Rhine graben. The minor faults *ff* hade towards F, and are inclined in the opposite direction to the beds. This arrangement is termed *antithetic.*

have been of a regional nature, or caused by the arching of the crustal block traversed by the graben.[1]

J. W. Gregory was the chief proponent of the idea that the

[1] Cloos, H., 'Zur Experimentellen Tektonik': *Die Naturwissenschaften*, Jhg. 18, 1930, pp. 714–47; *ibid.*, Jhg. 19, 1931, pp. 242–7. 'Zur Mechanik groszer Brüche und Gräben': *Cbl. für Min., Abt. B.*, 1932, pp. 273–86; 'Künstliche Gebirge': *Natur und Museum*, Vol. 59, 1929, pp. 225–72; *ibid.*, Vol. 60, 1930, pp. 258–69.

great system of fault troughs in East Africa consists also of true *rift valleys*, that is to say, the fault blocks are bounded by normal faults, which originated as a result of circumferential tension in the crust.[1] Wayland, however, maintains that they were caused by deep-seated lateral compression, and that the faults are reverse faults.[2] Bailey Willis[3] inclines to the view that vertical movements of uplift and depression were dominant, and that the effects of local compression observed in the field are subordinate. He suggests that the plateau-

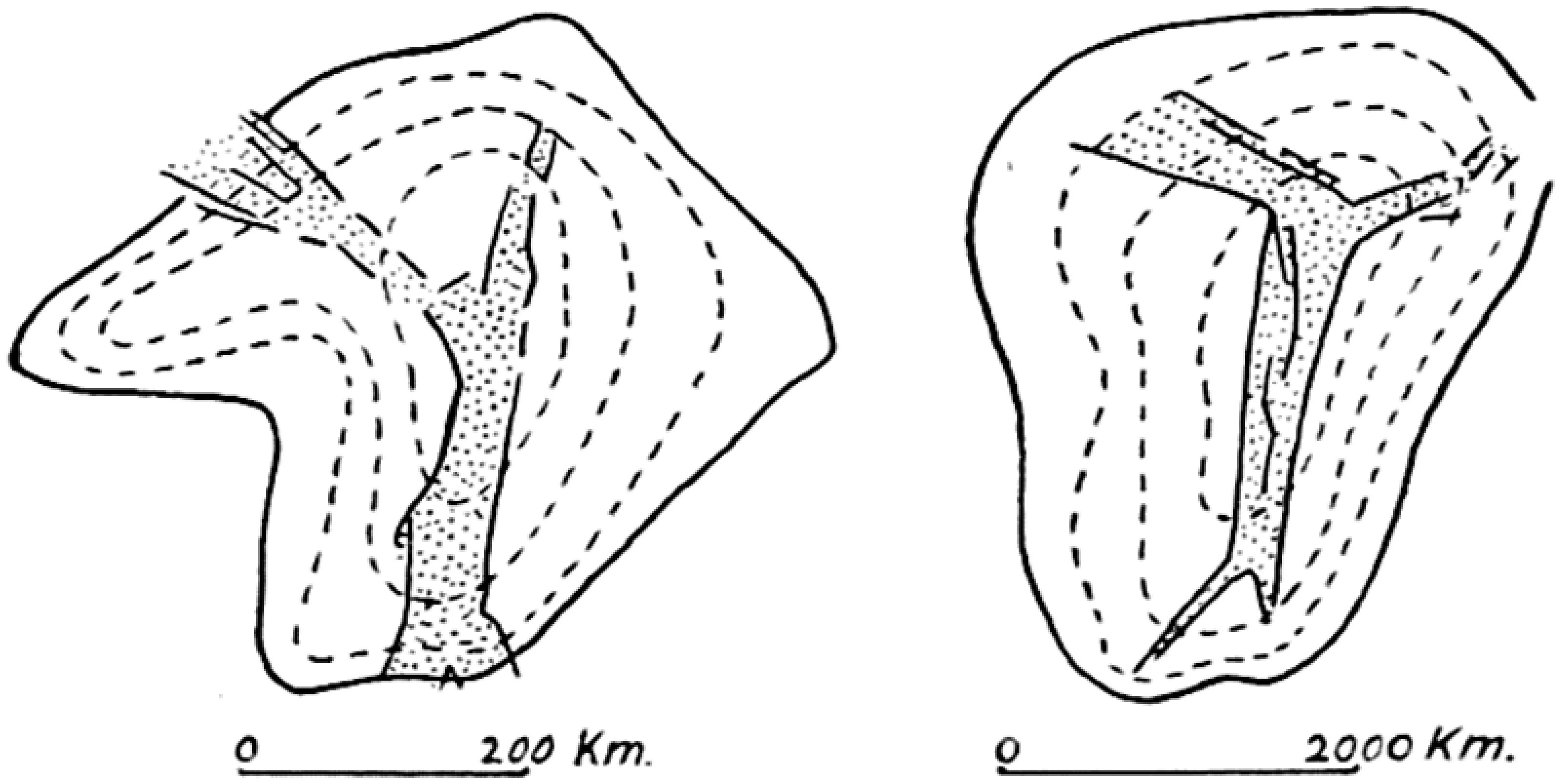


FIG. 42.—THE RHINE AND THE RED SEA RIFTS

(After H. Cloos)

The Rhenish (left) and Nubian-Arabian upwarps, with their associated grabens (dotted), reduced to the same width and with the Red Sea in reverse orientation north–south, to show analogous patterns.

uplifts may be caused by the expansion of formerly solid, though plastic, parts of the deep-seated asthenosphere, and ascribes the depressions to the removal of support from blocks adjacent to the plateaux, consequent upon escape of magma at the surface.

[1] Gregory, J. W., 'The African Rift Valleys': *Geogr. Journ.*, Vol. 56, 1920, pp. 13–47, 327–8; *The Rift Valleys and Geology of East Africa*: London, 1921.

[2] Wayland, E. J. 'Some Account of the Geology of the Lake Albert Rift Valley': *Geogr. Journ.*, Vol. 58, 1921, pp. 344–59.

[3] 'African Plateaus and Rift Valleys': *Carnegie Institute of Washington*, Pub. No. 490, 1936.

For the Dead Sea trough, Willis postulates boundary faults that are high-angle thrusts caused by lateral compression (see Fig. 43), a view which is essentially similar to Wayland's interpretation of the Lake Albert Rift. For valleys of this type Willis has coined the term *ramp valleys*.[1]

The compression theory does not explain such features as the zig-zag pattern of the fault troughs and the absence of strong folding in weak rocks lying in the troughs.[2] More recently, Cloos has demonstrated that the great grabens lie on

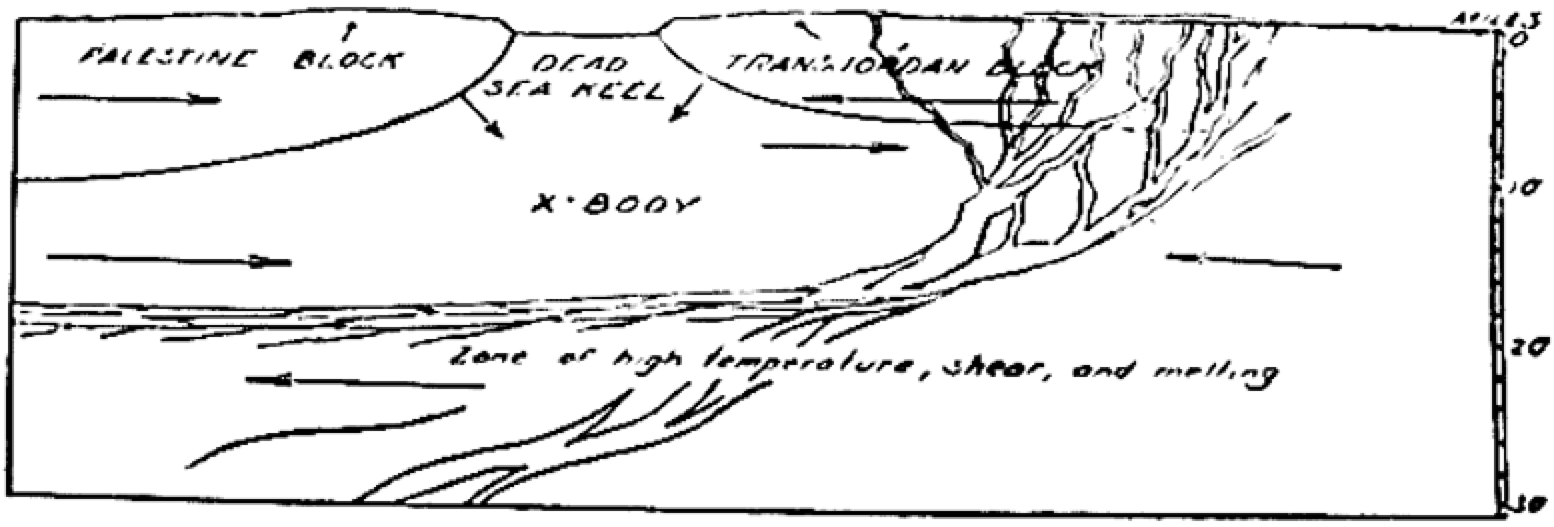


FIG. 43.—THE DEAD SEA TROUGH SHOWN AS A RAMP VALLEY

(After Willis, 1928)

The whole mass is regarded as under compression. The Palestine and Transjordan blocks have risen on the underlying ramps. The hypothetical X body has acted as a strut, and has been pushed towards the right. Melting has resulted from heating and shearing, and the liquid rock has been forced out as lava flows on the Transjordan block.

the crests of broad shield-shaped swells, and that certain similarities in pattern are shown in all, particularly in the bifurcation at the noses of the swells[3] (Fig. 42). The balance of evidence is clearly in favour of the tensional origin of grabens.

Well-known examples of block-faulted regions are the Basin

[1] Willis, B., 'The Dead Sea Problem, Rift or Ramp Valley': *Bull. Geol. Soc. Amer.*, Vol. 39, 1928, pp. 490–542.

[2] Faber, S., 'Fault Troughs': *Journ. Geol.*, Vol. 35, 1937, pp. 577–601. Bucher, W. H., *The Deformation of the Earth's Crust*: Princeton, 1933, pp. 325–39.

[3] Cloos, H., 'Hebung-Spaltung-Vulkanismus': *Geol. Rundsch.*, Vol. 30, 1939.

Ranges of Nevada and Utah,[1] and Fennoscandia,[2] but British examples of older block faulting, as in the Great Valley of Scotland and the coalfields of the Midlands, are known in greater detail.[3]

[1] Davis, W. M., 'Mountain Ranges of the Great Basin': *Mus. Comp. Zool. Harvard*, Vol. 42, 1933; 'The Basin Range Problem': *Proc. Nat. Acad. Sci.*, Vol. II, 1925, pp. 387–92. Gilbert, G. K., 'Studies of Basin Range Structure': *U.S. Geol. Surv.*, Prof. Paper 153, 1928, pp. 76–86.

[2] Sederholm, J. J., 'Weitere Mitteilungen über Bruchspalten mit besonderer Beziehung zur Geomorphologie von Fennoscandia': *Bull. Comm. Géol. Finlande*, No. 37, 1913, 66 pp. Asklund, B., 'Bruchspaltenbildungen in südöstlichen Östergötland': *Geol. Fören. Förh. Stockholm*, Vol. 45, 1923. Högbom, A. G., 'Zur Mechanik der Spaltenverwerfungen, etc.': *Bull. Geol. Inst. Upsala*, Vol. 13, 1916, pp. 391–408.

[3] Anderson, E. M., *The Dynamics of Faulting*: Edinburgh, 1942 (with references).

Chapter IV

FOLDS

1. *THE GEOMETRY OF FOLDS*

IN detailed structural studies a precise terminology for folded strata is necessary, and it is rather surprising to find that there is no general agreement among geologists as to the exact meanings of many of the terms used by them. In the following account of the geometry of folds most of the accepted modern usages are mentioned, and some indication is given as to which is considered preferable in cases where confusion might arise.

Types of Flexures

The various ways in which a particular bed or key-horizon may be folded are listed below.

These definitions arbitrarily assume the horizon as a reference plane, which is generally satisfactory but does not apply to folds pitching at 90° or more. In these it is preferable to define an anticline as a fold that includes older rocks in its core, and a syncline, younger rocks.

Difficulty may also be experienced in deciding on the structural nomenclature in block-faulted regions where there is strong drag or warping associated with the faults. In such regions, although fault-tectonics may be dominant, upward or downward flexures must be mapped as anticlines and synclines respectively.[1]

[1] Tromp, S. W., 'Blockfolding Phenomena in the Middle East': *Geol. en Mijnbouw*, Jhg. 11, 1949, pp. 273–8, has suggested that such folds be distinguished from compression-folds, but genesis is not implied in the use of the terms anticline and syncline, which refer simply to structures recognizable in the field.

An *anticline* is an upwardly convex flexure in which a given bed intersects the same horizontal plane in both limbs (Fig. 44, A).

A *syncline* is an upwardly concave flexure in which a given bed intersects the same horizontal plane in both limbs (Fig. 44, B).

An *anticlinal bend* [1] is an upwardly convex flexure in which one limb dips gently towards the apex (which see below), and the other limb dips more steeply away from it (Fig. 44, C).

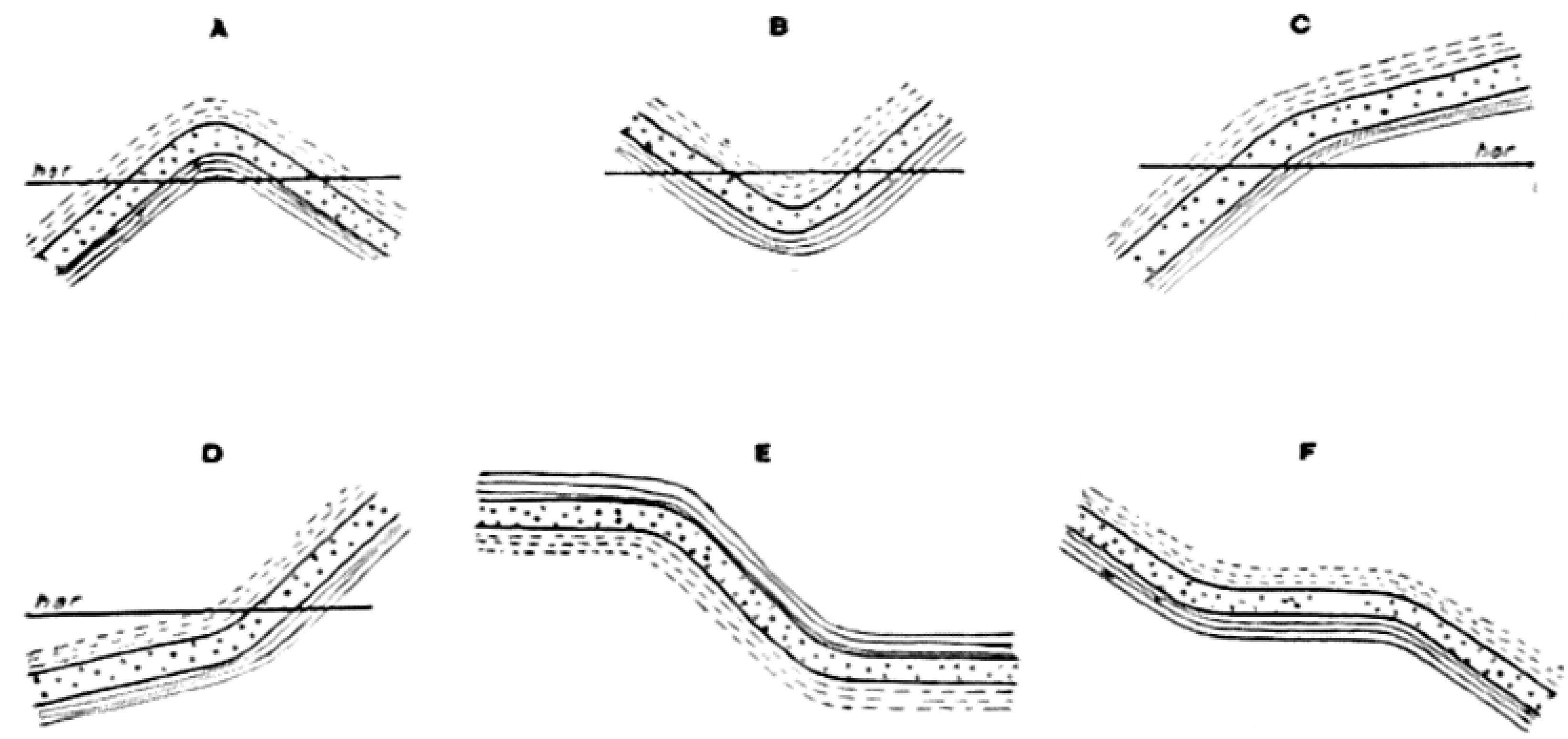


FIG. 44.—TYPES OF FLEXURES

A, symmetrical anticline; B, symmetrical syncline; C, anticlinal bend; D, synclinal bend; E, monocline; F, terrace ; *hor.*, horizon.

A *synclinal bend* is an upwardly concave flexure in which one limb dips relatively steeply towards the apex, and the other limb dips gently away from it (Fig. 44, D).

A *dome* is an upwardly convex flexure in which the dip is away from the centre in all directions (*quaquaversal dip*).

A *basin* is an upwardly concave flexure in which the dip is towards the centre from all directions. Domes and basins exhibit *periclinal structure*, or radial dip of the beds.

A *monocline* is a local steepening of an otherwise uniformly dipping or horizontal series, and is composed of an anticlinal bend above, followed by a synclinal bend at a lower level

[1] Busk, H. G., *Earth Flexures*: Cambridge, 1929.

(Fig. 44, E). In oil-field geology, however, the term is used for formations in which the dip is more or less uniform in one direction.

A *terrace* is a local flattening in an otherwise uniformly dipping series, and is composed of a synclinal bend above, followed by an anticlinal bend at a lower level (Fig. 44, F).

Anticlinal and synclinal structures on a large scale—of the order of miles in width—in which the limbs of the folds are themselves folded by minor plications, are called *anticlinoria* and *synclinoria* (singular, *anticlinorium*). The terms geanticline and geosyncline should not be used in this sense.

Description of Folds

Elements of Folds.[1]—The *crest* of an anticline is the line which joins the highest points of the fold, as defined by a particular bed (Fig. 45).

The *trough* of a syncline is the line which joins the lowest points of the fold as defined by a particular bed.

The *pitch* of a fold is the angle at which the crest of an anticline or the trough of a syncline dips at any point (see Fig. 69). Pitch may vary in direction and amount along the crest of a fold, and is measured in a vertical plane.

Some authorities[2] now use the word *plunge* in the sense given above for *pitch*, and have defined *rake* as the angle of inclination of a linear structure measured *in the plane of another structure* (e.g. lineation on a cleavage plane). If these usages are adopted the term pitch is discarded for folds.

The *crestal plane* of a fold is a surface so disposed within it that it contains the crests of successive beds within the fold (see Fig. 45).

The *axial plane* is defined by Busk[3] as a surface so disposed within the fold that any point upon that surface is equidistant from either limb of the fold, as defined by a particular bed.

[1] See Challinor, J., 'The Primary and Secondary Elements of a Fold': *Proc. Geol. Assoc.*, Vol. 56, 1945, pp. 82–8.

[2] *U.S. Geol. Surv.*, 'Explanatory Notes to a new List of Geological Map Symbols', 1948. The usages recommended in this Note have been adopted by several official organizations.

[3] *Earth Flexures*: p. 7.

The axial plane of a fold is not, however, a simple concept and it is dealt with more fully below.

Axial line is a term in general use, which may be used to designate the intersection of the axial plane of a fold with the ground surface, thus giving a useful term for map interpretation.

Crest line may then be defined in an analogous way.

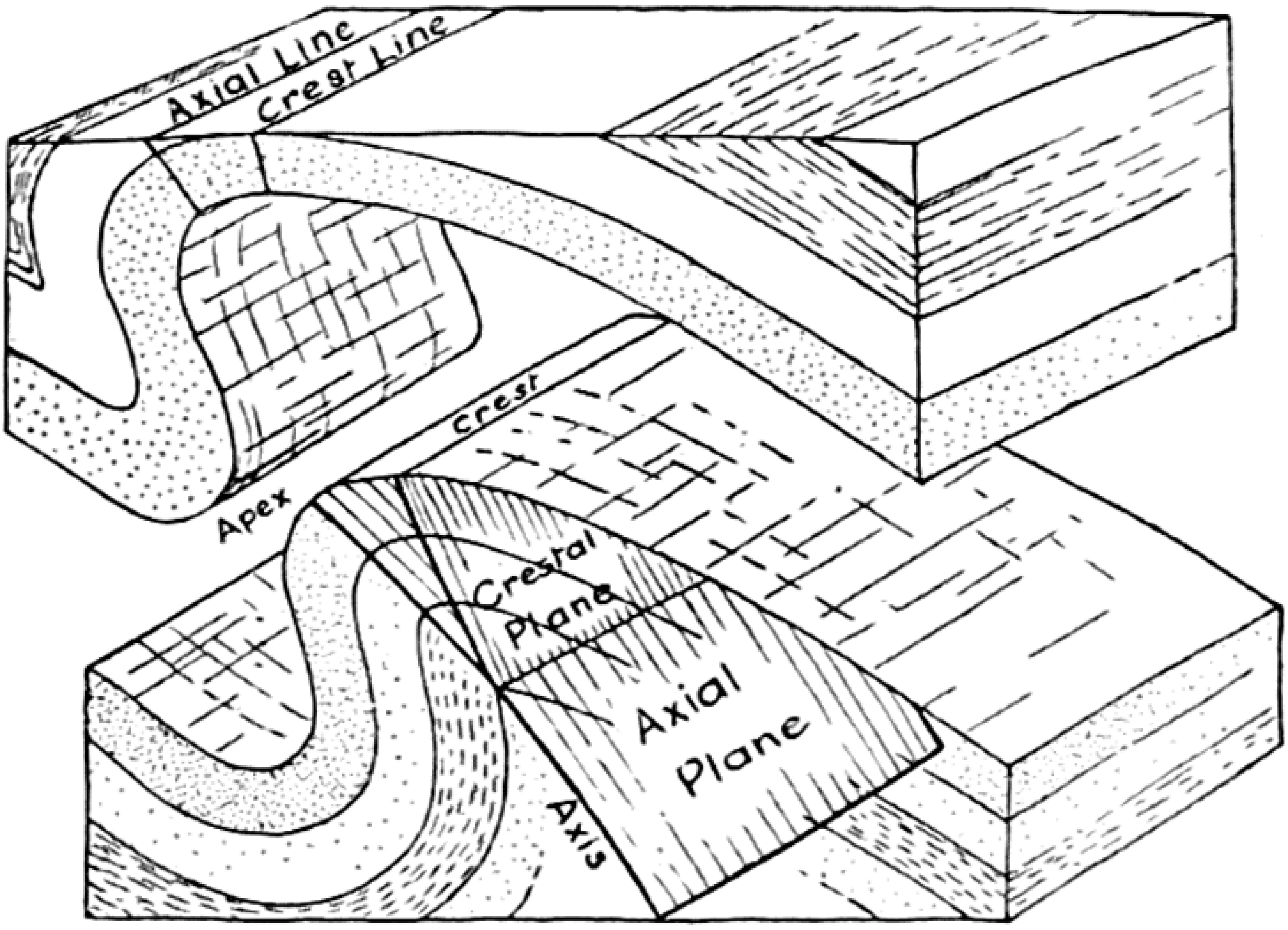


FIG. 45.—DISSECTED BLOCK DIAGRAM OF AN ASYMMETRICAL FOLD ILLUSTRATING THE TERMINOLOGY OF THE VARIOUS PARTS

The axial and crestal planes in the lower half of the diagram are shown as if seen through the beds.

Axis is a term used in several different senses. By B. and R. Willis[1] it is defined in the same way as crest and trough are defined above. Nevin[2] defines both axis and axial line as the 'median line of a fold, that is, the intersection of the axial plane and the bedding'. H. Cloos uses the term axis for the direction in which a fold trends, being 'diejenige Richtung, welche den Mantelinien einer Falte parallel ist'.[3] The term

[1] *Geologic Structures*: New York, 1934, p. 55.

[2] *Principles of Structural Geology*: New York, 1949, p. 41.

[3] *Einführung in die Geologie*: Berlin, 1936, p. 185.

is also used for the trace of the axial plane of a fold in a transverse section across it,[1] and it is suggested that this be adopted as the standard usage. It then only remains to decide upon a term for the intersection of the axial plane with a given bed, in order to complete the terminology for the various parts of a fold. The term *apex* is defined by Busk[2] as the line of inter-

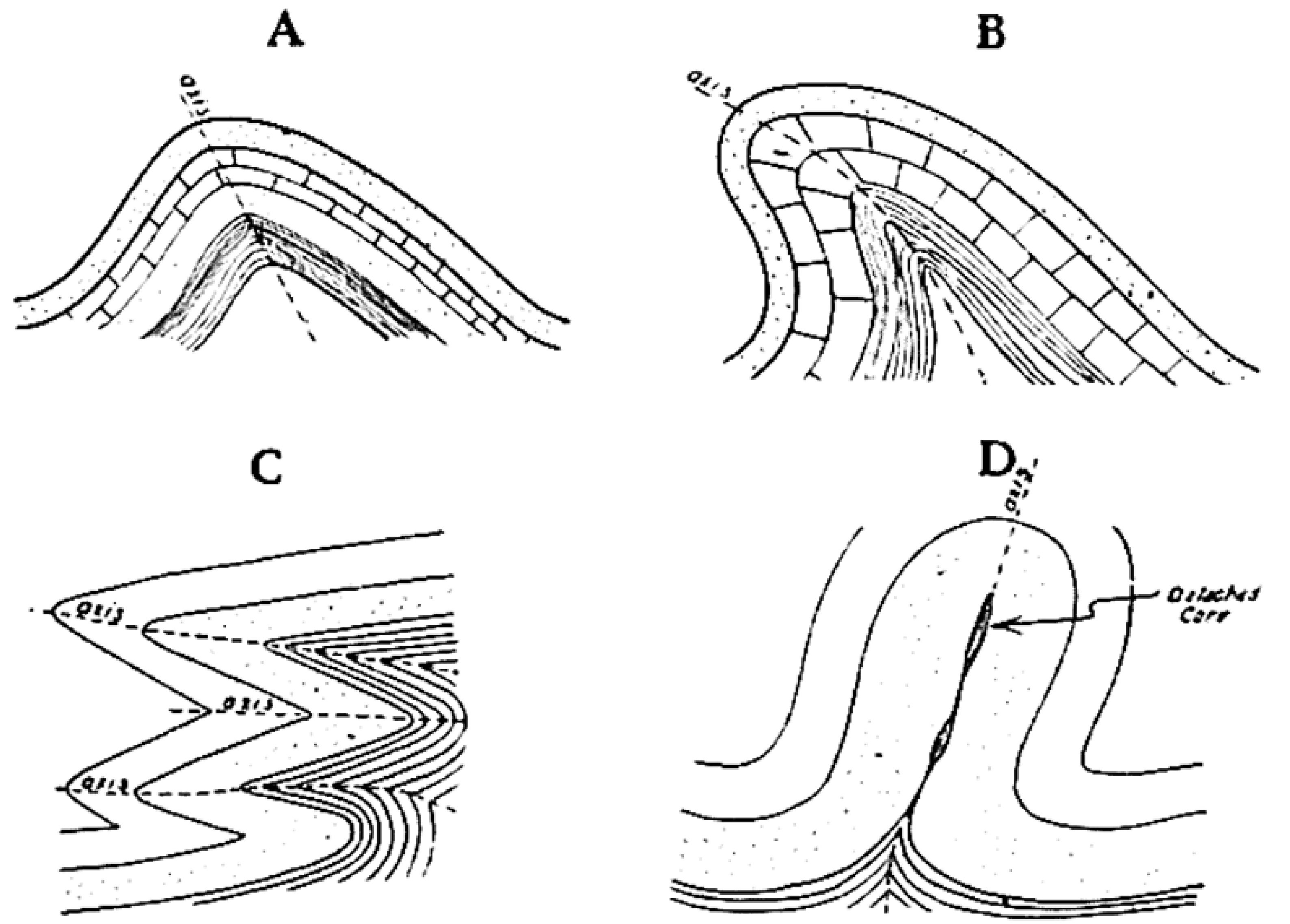


FIG. 46.—TYPES OF FOLDS

A, asymmetrical anticline (inclined);
B, ,, ,, (overturned);
C, recumbent zig-zag folds;
D, isoclinal fold with detached cores of shale.

section of the axial plane of a fold either with the ground surface or with a horizontal plane. Seeing that axial line has been used for some time, at least in Australia, in this sense, the *apex* is here defined as the line along which the axial plane intersects a given horizon in the fold. More directly stated, the apex is the point, in a cross-section, where the rate of change of

[1] Dana, J. D., *Manual of Geology*: New York, 1880, pp. 93, 95.
[2] *Earth Flexures*: Cambridge, 1929, p. 7.

dip is greatest, and the fold is, therefore, most acute. Along the fold, the apex is a line comprising such points.

The nomenclature herein proposed permits of the description of a fold with reference to each bed within it, this being often required in economic work. An illustration of the advocated usage is given in Fig. 45.

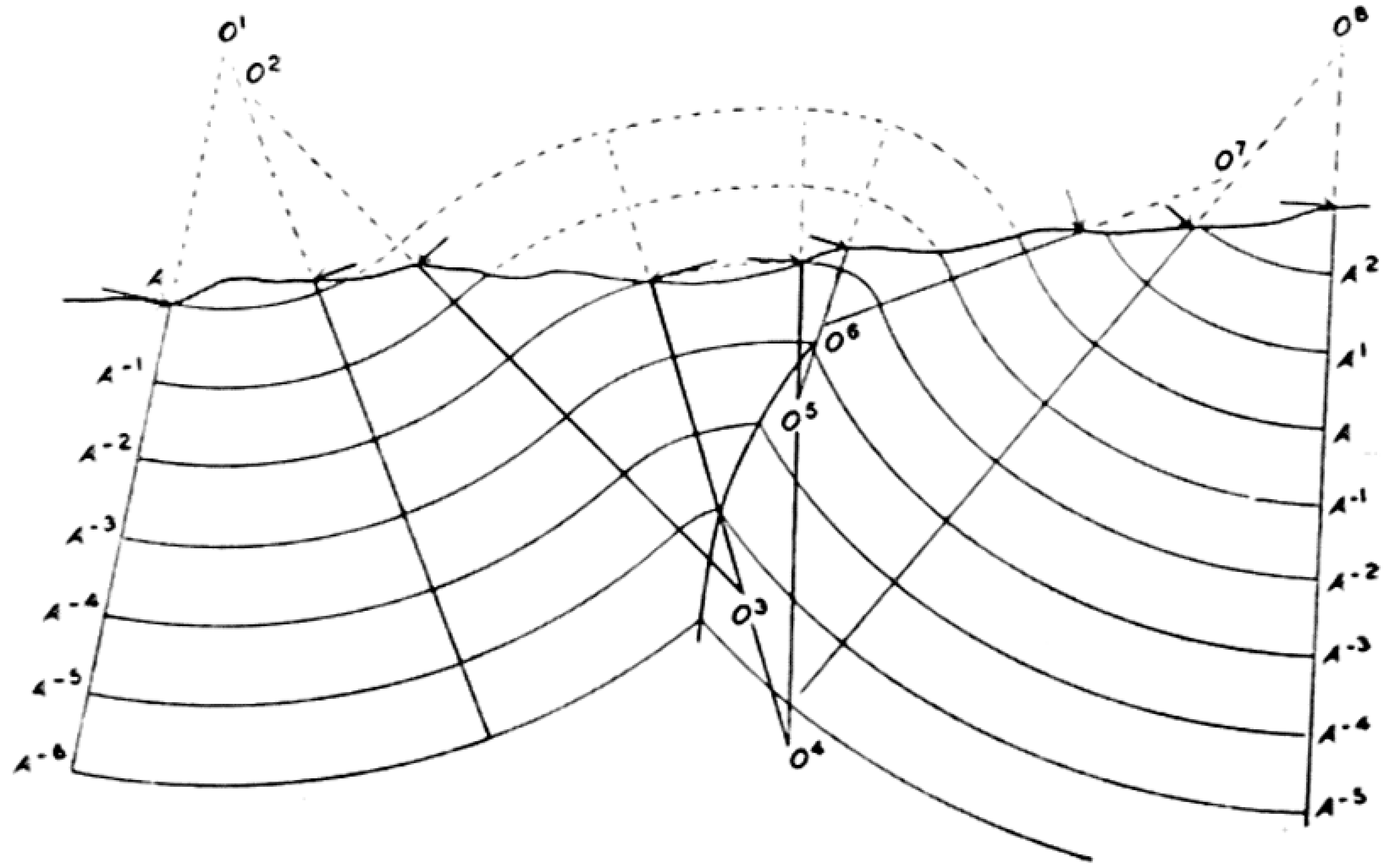


FIG. 47.—AN ASYMMETRICAL FOLD OF THE PARALLEL TYPE, REPRESENTED BY STRATIFICATION PLANES A, A^1, A^{-1}, &C., CONSTRUCTED AT UNIT DISTANCE APART

(After Busk, *Earth Flexures*)

Between observed dips at outcrop the beds are assumed to have the form of circular arcs whose centres or origins—O, O^1, &c., are given by the intersections of radii drawn at right angles to the dip arrows.

Symmetry and Attitude of Folds.—Various terms are used to describe the attitude of folded beds as seen in cross-section at any one place. A fold is usually said to be *symmetrical* if the dip of any bed, measured at the points of intersection of the bed with a given horizontal plane, is the same in each limb (see Fig. 44, A). According to this definition the axial plane must be vertical, but Nevin and others[1] would describe a fold as

[1] Nevin, C. M., *Principles of Structural Geology*: New York, 1931, pp. 36–8. Stočes, B., and C. H. White, *Structural Geology*: London, 1935, p. 116.

symmetrical if the beds in either limb dip at similar angles with the axial plane, even if this be inclined. Stočes and White[1] would restrict the use of the term to folds which are not only symmetrical in this sense but in which the limbs are, in addition, of equal length. Both the latter definitions are contrary to general modern usage.

In *asymmetrical folds* the dips of the two limbs at the intersection of any bed with a given horizontal plane are different. If the axial plane is inclined, but the steeper limb is not overturned, the fold is said to be *inclined*; if the steep limb is overturned, the fold is an *overfold*; if the axial plane becomes nearly horizontal, the fold is *recumbent*, and if the two limbs are approximately parallel, the fold is *isoclinal*.

Relationships of Strata in Folds

Parallel Folds.—Consideration of the attitude of the successive beds in a series of folded sediments led van Hise[2] to distinguish between two distinct types of arrangement, which he called *concentric* or *parallel folding* on the one hand, and *similar folding* on the other.

In parallel folding, the strata are bent into parallel curves, and retain a constant thickness throughout. Thus, the dips decrease both upwards and downwards, owing to the progressive increase in the radii of curvature of the beds in each sector, with increasing distance from each centre of curvature, or *origin*.[3] Such folds, as may be seen by bending a pack of cards held firmly at one end, can take place only with the accompaniment of differential sliding movements between the beds. The uppermost of any two beds in an anticline moves over the lower in the direction of the apex of the anticline. In each bed, internal strain takes place as illustrated in connexion with bending (Fig. 22).

[1] *Structural Geology*: London, 1935, p. 116.

[2] van Hise, C. R., 'Principles of North American Pre-Cambrian Geology': *16th Ann. Rept. U.S. Geol. Surv.*, Pt. 1, 1896, pp. 581–843.

[3] Since a smooth curve may be resolved into a number of tangential arcs of circles, it is permissible to regard the beds in a small sector of a concentric fold as having the form of concentric circular arcs, so that we may speak of a centre of curvature for each segment.

Idealized diagrams (Fig. 48, A) show parallel folds dying out upwards and downwards from a median zone of curvilinear folds in which zone the shortening of the section is apparently a maximum. Since, however, the fold axes remain vertical the shortening is, in fact, constant, and in such folding the upper and lower beds in which the folds are cuspate, must have thickened. In nature, the conditions for parallel curvilinear folding are satisfied for relatively strong ('competent') beds that are either interbedded with or overlie relatively weak, plastic ('incompetent') beds, which latter yield by complex

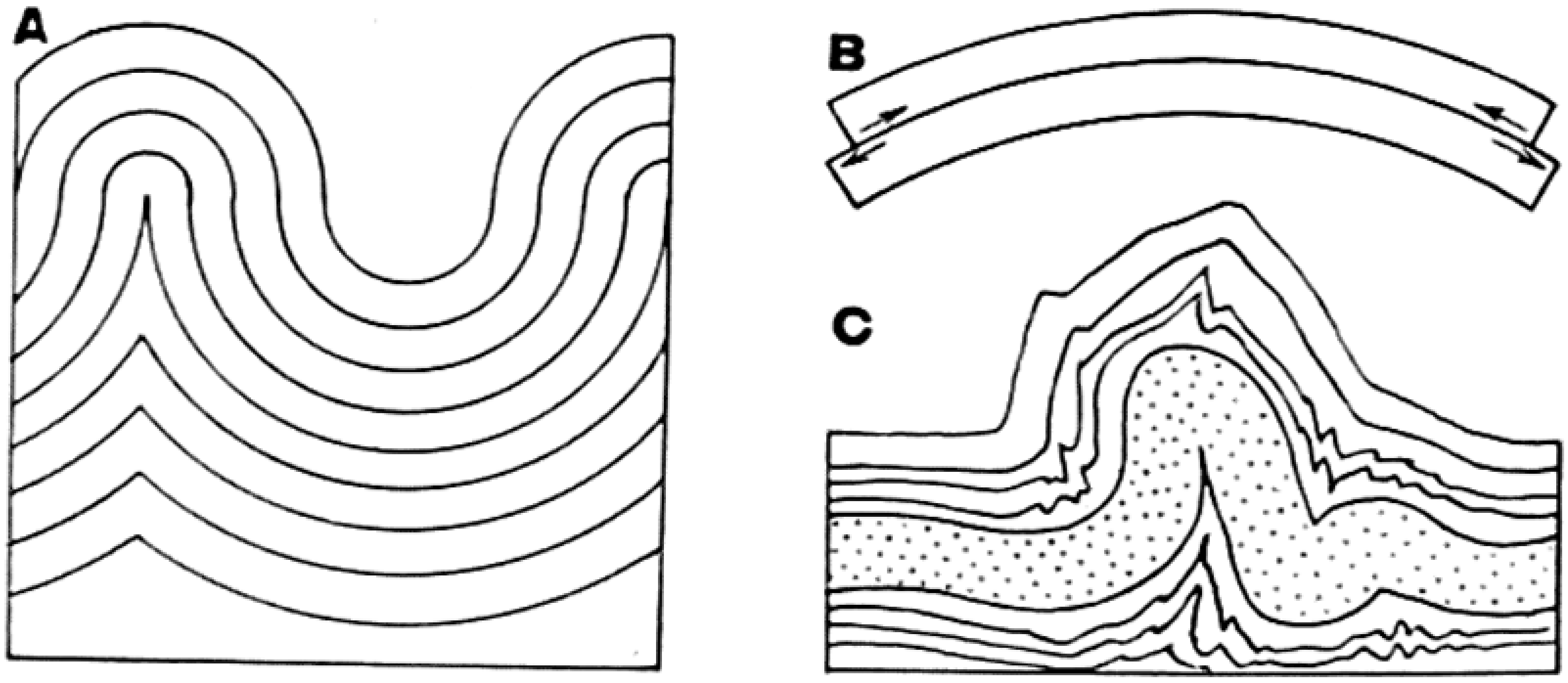


FIG. 48.—PARALLEL FOLDS

A. Idealized upright parallel folds, showing sinusoidal and cuspate zones, and decrease in fold intensity in depth.
B. Parallel folding of two beds, showing bedding-slip.
C. Parallel folding in one formation (dotted), with disharmonic relationships in under- and overlying beds.

folding or shearing. These are the conditions in Willis' classic experiments[1] and the implication is that there construction of parallel folds in section-drawing should normally cease at the contact of the competent beds with the enclosing incompetent beds, which yield in other ways (Fig. 48, B). In a thick series of strata of overall competence, the required shortening above the zone of curvilinear folding may take place by thrust faulting. These various effects are well shown in the Juras (Fig. 33).

[1] Willis, B., 'The Mechanics of Appalachian Structure': *13th Ann. Rept. U.S. Geol. Surv.*, Pt. 2, 1893, pp. 217–81.

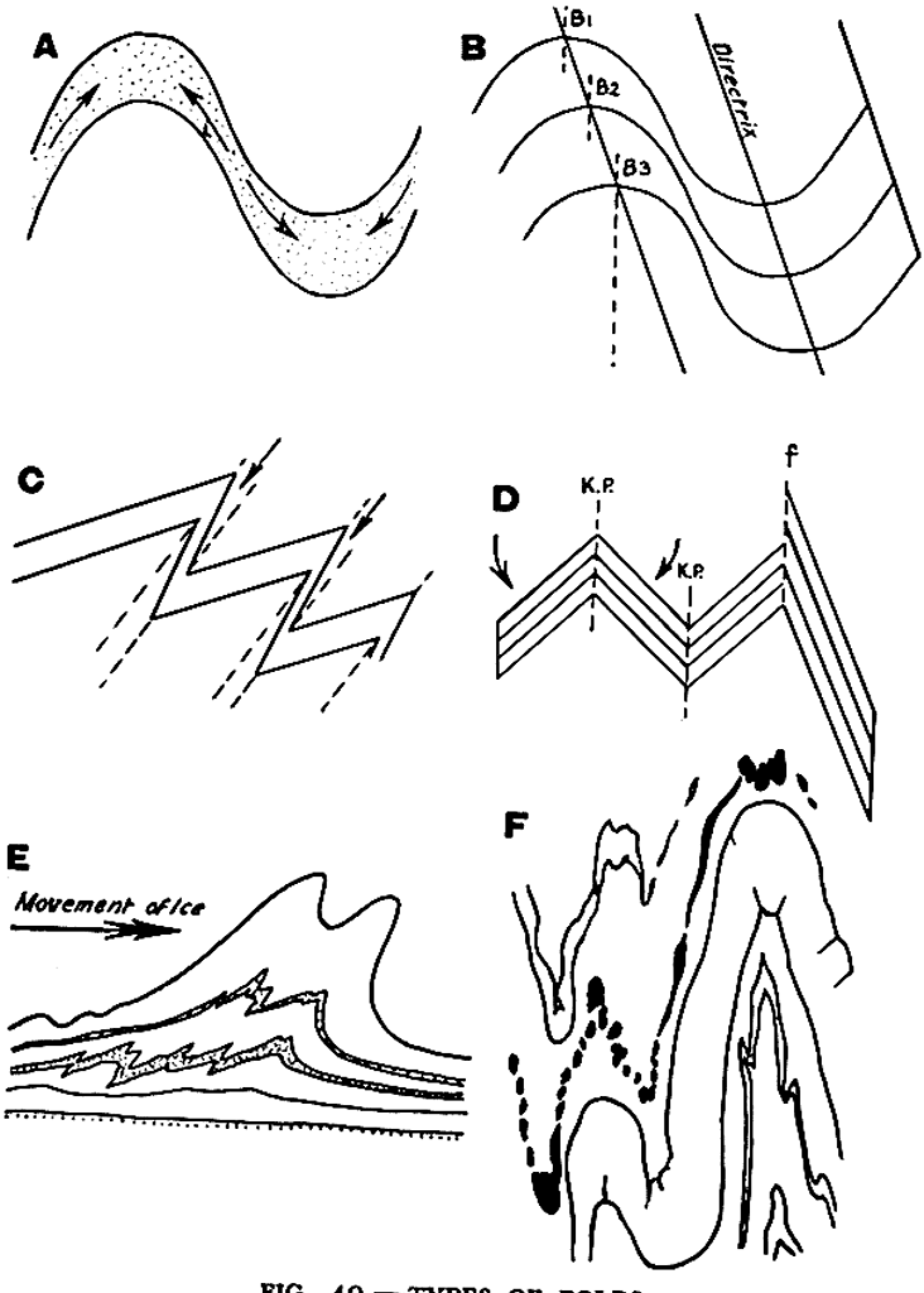


FIG. 49.—TYPES OF FOLDS

A. Upright symmetrical similar fold, with arrows indicating supposed flowage from the limbs.

B. Inclined symmetrical fold, sinusoidal shape, showing differing bisectrices B1, 2, 3 in each bed, and the common direction of generation (directrix).

C. Chevron or zig-zag folds due to shearing of a uniformly dipping bed.

D. Chevron or zig-zag folds due to rotation about knick-planes (K.P.) and axial faulting (f) where fold is asymmetrical.

E. Generative fold in Tertiary brown coal, Germany, due to drag of Pleistocene ice sheet (after Seidl).

F. Disjunctive fold (in black bed) shear and buckle-folding in other beds, in hematite-quartzite (hand-specimen).

Similar Folds.—In similar folding on the other hand, the strata are bent into similar curves, and the folds maintain their identity in a vertical direction (Fig. 50). In this case, it will be seen (Fig. 49, B) that the beds do not retain their original thickness throughout, but the limbs of the folds are thinner than the apices. This may take place as a result of flowage of material from the limbs towards the apices, but other causes more probably operated in most instances, for in a similar fold that persists for thousands of feet in depth, this concept would imply that all plastic beds flowed to the same extent and that all more rigid beds took the same fold-form despite physical differences between them. Consideration of asymmetrical similar folds such as are formed as an early stage of stretch-thrusts demonstrates, however, that the geometry of similar folds is satisfied by shearing parallel to one direction, which may be designated the *directrix*. In such folds the original thickness of the beds is shown at the apex, and the limbs are stretched to various degrees.

Zig-zag Folds.—Zig-zag or chevron folds afford a special case exhibiting certain features of both parallel and

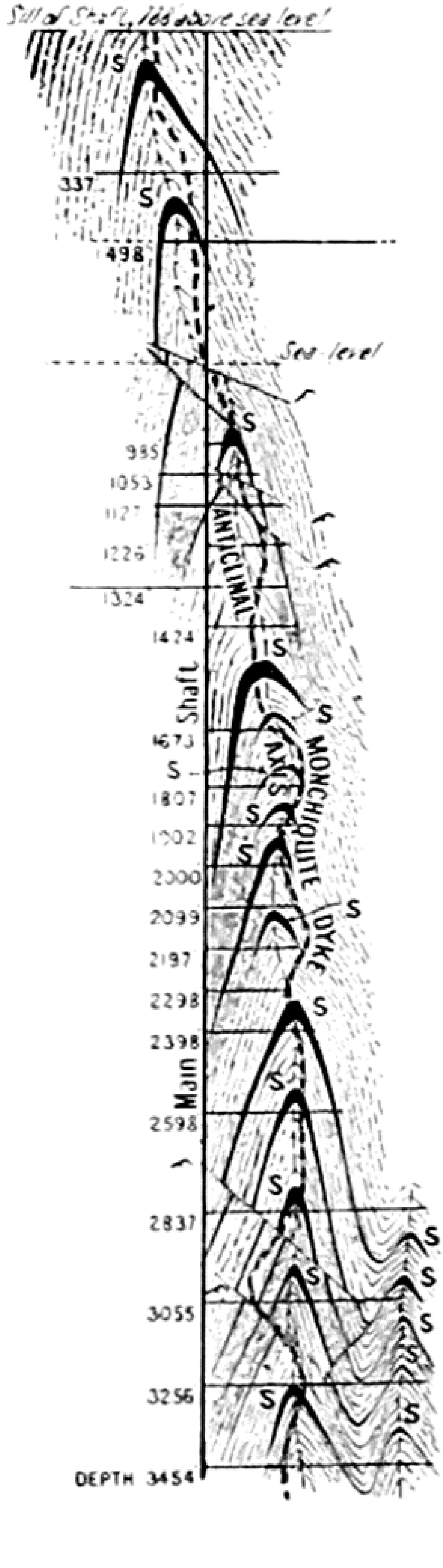


FIG. 50.—SECTION AT THE GREAT EXTENDED HUSTLERS SHAFT, BENDIGO, TO ILLUSTRATE SIMILAR FOLDING

(From *Economic Geology and Mineral Resources of Victoria*, by H. Herman: *British Association Handbook to Victoria*, 1914)

Note the persistence of the fold in depth, and the saddle reefs (S) which serve to indicate the thickening of the whole formation at the fold axis. A Tertiary monchiquite dyke follows the surface of weakness afforded by the axial plane of the fold.

similar types, since the bedding planes remain parallel in the limbs but the fold-form persists in depth. Two chief types may be recognized, the first involving the shearing of a uniformly dipping bed (Fig. 49, C), the second, rotation of the beds about the axial planes, as with two packs of cards pushed together and each turned so as to make a ridge. This is known as *knicking*, and the planes about which rotation occurs, *knick-planes*. Unless the rotation of each limb is the same, the steeper-dipping limb must rise above the gentler at the knick-plane, which becomes a fault (Fig. 49, D).

In this mechanism, bedding slip takes place as in parallel folding, but unless the beds actually break at the fold axes special readjustments are required there since the beds are locally bent. Chevron folds may combine knicking and shearing in their formation, the intimate shearing of relatively plastic strata giving rise to cleavage.

Supratenuous Folds.—A fold which shows a 'thinning of formations upwards, above the crest of the fold',[1] is termed *supratenuous*. As we have seen in the discussion of folding of the

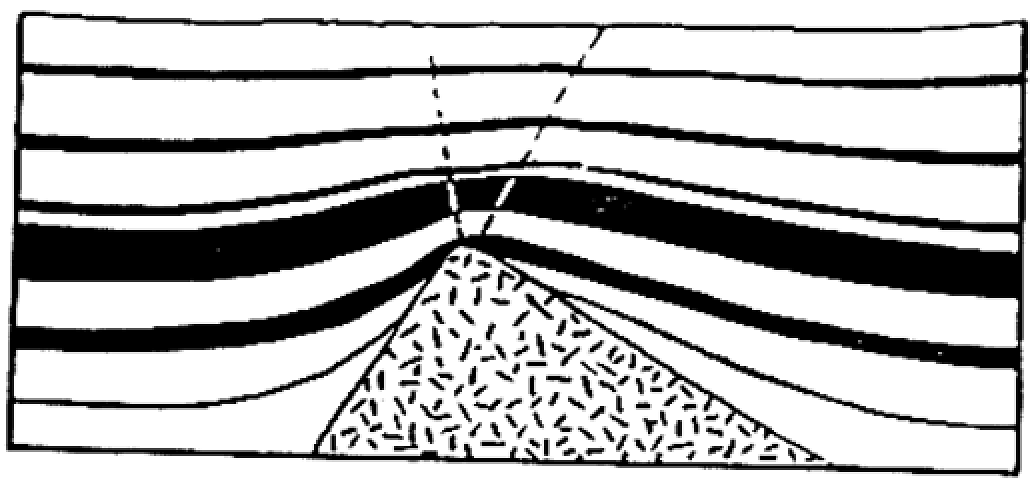


FIG. 51.—SUPRATENUOUS FOLD PRODUCED IN AN EXPERIMENT WITH SAND AND CLAY DEPOSITED IN SUCCESSIVE LAYERS OVER A PROMINENCE

(After Nevin, *Principles of Structural Geology*)

Note the inclination of the crestal plane towards the steeper side of the 'buried hill', and towards the steeper dipping limb of the fold.

Plains type, this feature may arise from the upward thrust of buried hills on superincumbent sediments, or from differential compaction of sediments lying on an irregular basement. It

[1] Nevin, C. M., *Principles of Structural Geology*: New York, 1931, p. 47.

may also develop if folding and sedimentation are contemporaneous, the formations then being thicker in the synclinal troughs than on the anticlines, when originally laid down. Asymmetry of the buried hills causes the anticlines to be asymmetrical, and experimental studies[1] have shown that the inclination of the crestal plane of an anticline is controlled by the slopes of the underlying buried hill (Fig. 51). The crestal plane dips towards the steeper slope of the hill and also towards the steeper limb of the fold, down to the point where the axial and crestal planes meet—at the summit of the buried hill.

Generative Folds.—The term *generative* is here applied to folds that increase in amplitude in successive beds with the accompaniment of an increase in stratigraphic thickness towards the axial region. Such folds involve the plastic mass-deformation of a group of beds, but the effects may be localized, as in a fold with one stretched limb in which the stretching increases to a maximum and then decreases again. Generative folds are common in coal basins, and in Tertiary brown coals in Germany they are demonstrably formed by plastic deformation due to the drag of Pleistocene ice-sheets (Fig. 49, E).

Disharmonic Folds.—Where an abrupt change in fold geometry takes place in passing from one bed to another the folding is said to be *disharmonic*. This is commonly found where relatively plastic strata are interbedded with more rigid beds (see e.g. drag folding, pp. 97–9), and is the rule rather than the exception in strongly deformed rocks.

Disjunctive Folds.—Where relatively brittle beds are interbedded with plastic rocks that are strongly deformed by flowage, the brittle beds may fracture and the parts become separated, while retaining an overall fold pattern. This may occur on a large scale as well as in small exposures. Disjunctive effects are common in soft-rock folding, as in slump-masses; in normal lithified sediments they are often seen in limestones interbedded with shale, and in salt tectonics; again, they occur in gneisses and schists (Fig. 49, F).

[1] Nevin, C. M., and R. E. Sherrill, 'Studies in Differential Compaction': *Bull. Amer. Assoc. Petrol. Geologists,* Vol. 13, 1929, pp. 1–22.

Diapiric Folds.—In *diapiric folds* (*piercement folds*, *plis diapirs*, *Injektivfalten*), mobile beds are actually injected through overlying strata at anticlinal axes (see Fig. 52). This commonly occurs with salt deposits (see pp. 94–6), and is also often shown by shales in regions of strong folding.

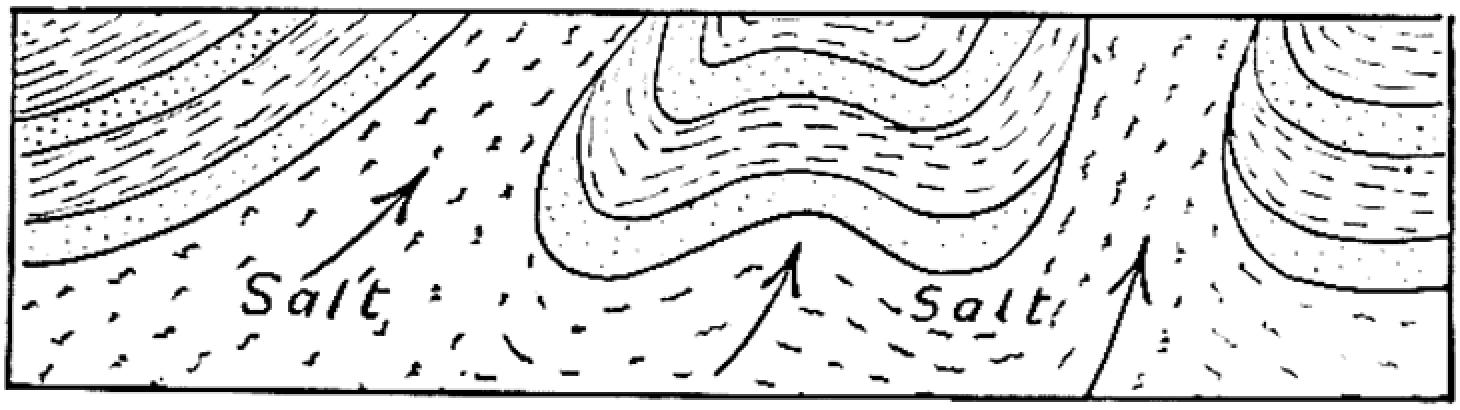


FIG. 52.—PIERCEMENT FOLDS, SHOWING SALT DEPOSITS INJECTED UPWARDS IN THE DIRECTION OF THE ARROWS, AND CUTTING THROUGH ANTICLINES IN THE OVERLYING BEDS

(After Pustowka, *Neues Jahrb. f. Min.*, B.B., 61, Abt. B, 1929, p. 373)

Axial and Apical Planes

The definition of the axial plane of a fold previously given (p. 77) implies that this plane (or surface[1]) bisects the fold, a usage adopted by most authors. This definition, however, is not applicable to asymmetrical similar folds, in which each bedding plane has a different bisectrix. In such folds it is the plane connecting the apices of the beds in the fold that is significant. This apical plane is parallel to the directrix of the fold (Fig. 49, B). In a parallel fold that possesses an apex the apical plane and the axial plane are identical, but sinusoidal parallel folds have no apices and no apical planes, and the axial plane is then identical with the crestal plane.

In many parallel folds, it is seen that subsidiary flexures (anticlinal or synclinal bends) occur on one or on both limbs. The apical planes of these flexures are of fundamental importance in folding, as may be seen in Willis' experiments[2]

[1] Challinor has proposed the use of *axial surface* for axial plane, but the word plane is often used for curved surfaces in geology (e.g. shear plane) and is retained in the present context.

[2] Willis, B., 'The Mechanics of Appalachian Structure': *13th Ann. Rept. U.S. Geol. Surv.*, Pt. 2, 1893, Pls. 79, 81, 82, 86, 93.

and is inferred from detailed sections of actual folds available from mines. The apical planes of the lateral flexures are knick-planes about which rotation of the beds goes on as the folds become more strongly developed. Stages and types of folding

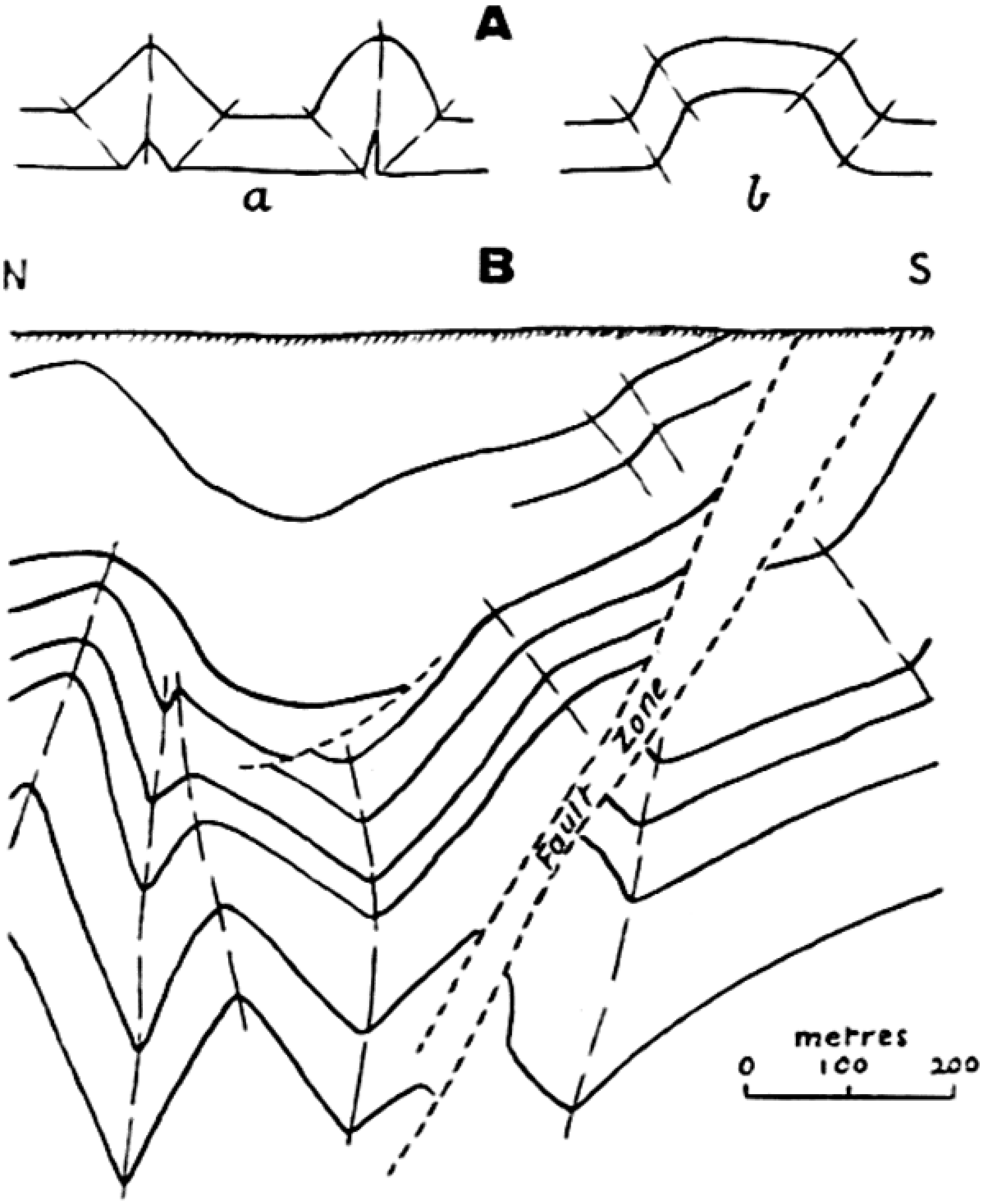


FIG. 53.—KNICK-PLANES IN FOLDING

A. Median and lateral knick-planes (*a*) and pairs of lateral knick-planes (*b*) in experimental folds (after Willis).
B. Section through the Bochum Basin, Ruhr coal-field, showing increasing fold intensity in depth, and 'axial planes' regarded as knick-planes (after Böttcher).

by this mechanism are shown in Fig. 53. It will be seen that in rectangular folds (*Kofferfalten*) the lateral apical planes are of greater significance than the 'axial plane' of the main fold itself, although this remains the direction of general upward movement. Lateral apical planes may develop, after strong

folding, into thrust faults. Where the axial plane persists and a fold becomes isoclinal, plastic beds are squeezed out of its core (downwards in an anticline), residuals remaining as *detached cores* (Fig. 46, D). Thus the axial and apical planes are of considerable genetic significance as well as being important in the geometry of folds.

2. *MECHANICS OF FORMATION OF INDIVIDUAL FOLDS*

In order to understand the relationship of minor structures to folding, it is necessary to consider the genesis of individual folds rather than of zones of folded rocks.[1]

The chief types of folds and fold-like structures in rocks may be classified according to their mode of origin as *buckle folds*, *bending folds*, *slip* or *shear folds*, and *flow folds*, but a particular fold may show the characters of more than one of these types.

Buckle Folds.—A sheet of material is said to be buckled when it is thrown into a double flexure by compression acting in the plane of the sheet, the flexures arising where irregularities in the sheet cause resolutes of the stress to act normal to it (upwards or downwards in an originally horizontal sheet). The stresses in the sheet may be caused by the end-on push of forces applied to its ends, in which case the sheet transmits the stress, or they may be due to forces applied to all parts of the sheet simultaneously, as with the pull of gravity on an inclined sheet. In all examples of buckling the length of the folded section is shortened and the fold-geometry is determined, under

[1] The following are important works relating to this subject: Tromp, S. W., *On the Mechanism of the Geological Undulation Phenomena in General and of Folding in Particular*: Leiden, 1937. Sander, B., *Gefügekunde der Gesteine*: Vienna, 1930, pp. 243–75. Schmidt, W., 'Gesteinsumformung': *Denksch. Naturhist. Mus. Wien*, Bd. 3, 1925; *Tektonik und Verformungslehre*: Berlin, 1932. Cloos, H., *Einführung in die Geologie*: Berlin, 1936, pp. 200–12. Willis, B., 'The Mechanics of Appalachian Structure': *13th Ann. Rept. U.S. Geol. Surv.*, Pt. 2, 1893, pp. 217–81. De Sitter, L. U., 'Notes on the Mechanism of Folding': *Leidsche Geol. Med.*, Deel 8, 1936, pp. 161–8. Seidl, E., 'Formen der technischen Mechanik und ihre Anwendung auf Geologie': Bd. 5, Krümmungsformen, 1934.

given conditions, by the relative plasticity of the formations present. Beds that do not exhibit mass-flowage, although they may yield sufficiently to flex, determine the general fold-pattern and may be used to estimate the extent of shortening involved in the folding. More plastic beds may yield by flow folding or shear folding.

Bending Folds.—Folds which are caused by vertically acting forces of different intensity in different places may be called *bending folds.*[1] They do not involve lateral shortening of the section of folded rocks such as characterizes buckle folds, but the arching of beds by push from beneath increases the surface area of the deformed rocks. The stretching may take place by flowage from the anticlinal crest towards the limbs of the fold, giving a supratenuous fold, or by normal faulting[2] (see Fig. 54).

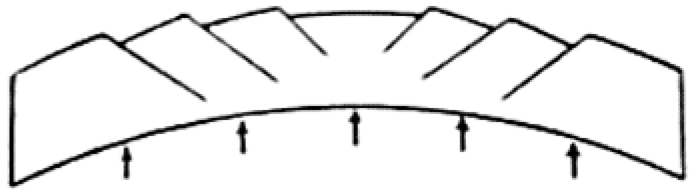


FIG. 54.—A BENDING FOLD PRODUCED BY PUSH FROM BENEATH
(After Cloos, *Einführung in die Geologie*)

The stretching is shown as taking place by the development of normal faults. In other cases flowage from the crest towards the limbs takes place.

Many large crustal warps are undoubtedly of this origin, and it is also important in folding of the Plains type[3] and in *Bruchfalten*. The deformation in these types is more localized along basement fault lines, stretching of the intervening portions not being necessarily involved.

Slip or Shear Folds (Scherfalten).—Differential movements along closely spaced shearing planes in rocks may cause originally plane structures to become curved so that they resemble true flexures. The development of such *slip* or *shear folds*, as

[1] Robinson, W. I., 'Folds resulting from Vertically Acting Forces': *Journ. Geol.*, Vol. 31, 1923, pp. 336–43.

[2] Cloos, H., 'Über Biegungsbrüche und selektive Zerlegung': *Geol. Rundschau*, Vol. 24, 1933, pp. 203–19.

[3] Also in the Uinta type of Powell and Mellard Reade (*The Origin of Mountain Ranges*: London, 1886), defined as a flat arch with marginal monoclinal flexures.

they are termed,[1] may be illustrated by ruling a straight line across the edges of a pack of cards, and then pushing in the end of the pack. Becker[2] was the first to recognize the significance of differential movements along shearing planes in producing curved structures in rocks, but some of these structures would not normally be termed folds.

Shearing planes sufficiently closely spaced to permit shear folding to develop must constitute cleavage (or schistosity) and

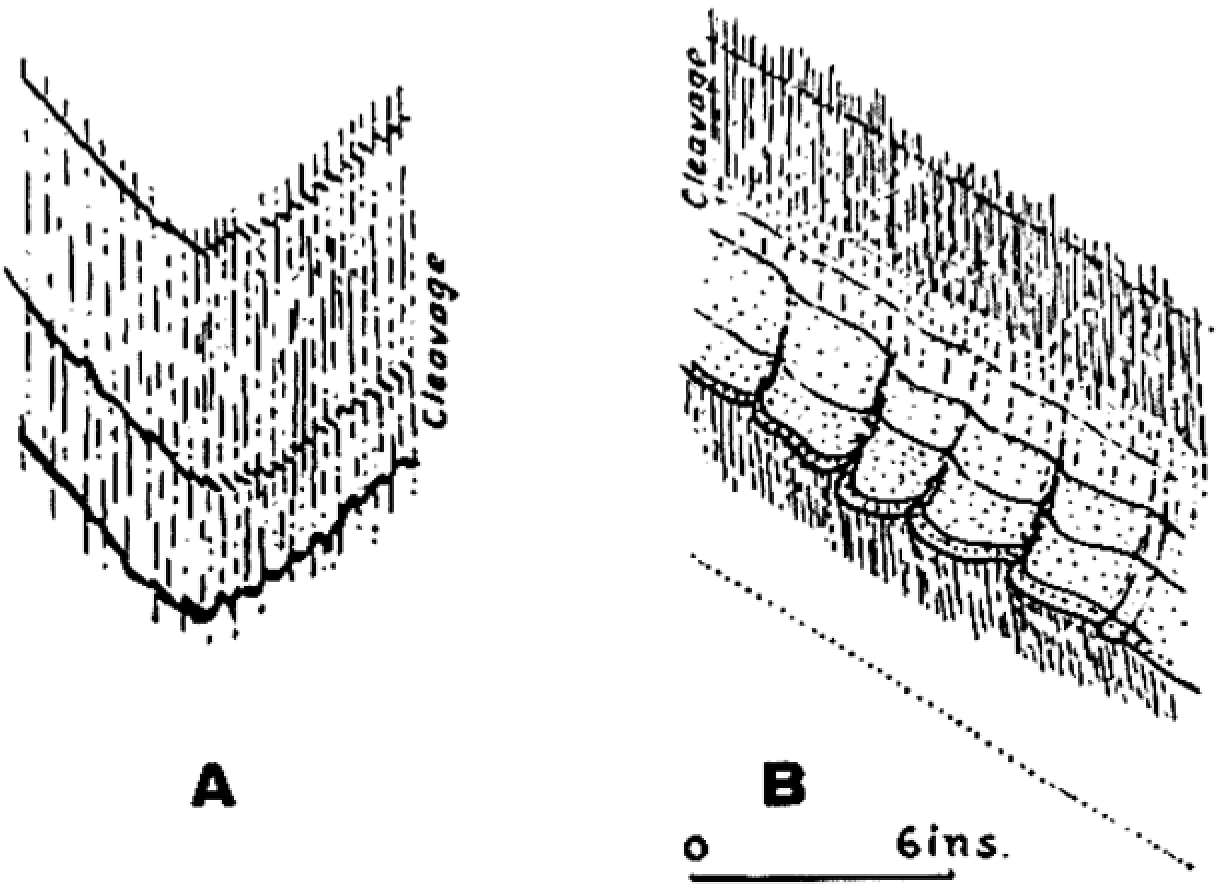


FIG. 55.—FOLDING OF SANDY LAMINAE IN SLATES

A. Shear-folding in very thin laminae (xl).

B. Folding of base of thin graded sandstone in slate, grading to smaller undulations. Note absence of folding in extremely thin laminae, in which shear is distributed, rather than localized in zones.

it is demonstrable in many slates that slipping along the cleavage, as measured by the displacement of thin sandy or coloured laminae, is sufficient to account wholly or partially for the form of folds (Fig. 55, A). Schmidt[3] has recognized the significance in folding of the spacing of zones of shearing, the

[1] Sander, B., *Gefügekunde der Gesteine*: Vienna, 1930, pp. 243–51. Knopf, E. F., and E. Ingerson, 'Structural Petrology': *Mem. Geol. Soc. Amer.*, No. 6, 1938, p. 157.

[2] Becker, G. F., 'Geology of the Comstock Lode': *U.S. Geol. Surv.*, Mon. No. 3, 1882.

[3] Schmidt, W., *Tektonik und Verformungslehre*: Berlin, 1932.

slices of rock between which are known as *Gleitbretter*. Where the shear zones are widely spaced the folds are larger, closer shearing giving smaller folds. This is also demonstrable on a small scale in slates, where shear zones are more widely spaced in sandy layers, which become folded while the thin laminae are intimately cleaved (Fig. 55, B).

In shear folds the thickness of a given bed measured along the shearing planes remains constant, but the apices of folds appear thicker than the limbs if the measurement is made at right angles to the bedding planes. In the absence of space-data relating to deformation, cross-sections of highly cleaved or schistose rocks are often better constructed on this principle rather than with the retention of constant stratigraphic thicknesses.[1]

Flow Folds.—The convolutions of different-coloured bands in moving liquids, well shown by stirring an oil film on water, serve to illustrate what is termed *flow folding* in rocks.

In flow folding the beds offer so little resistance to deformation that they assume any shape impressed upon them by surrounding more rigid rocks or by the general stress-pattern of the deformed zone. Individual strata are thickened or stretched accordingly, and the folds cannot be 'straightened out' to estimate the shortening involved. Flow folds may form in incompetent beds among other strata that yield by buckling, but thin competent beds involved in strong flow folds may be disrupted, the fold geometry being then determined by the incompetent strata.[2]

Some of the complex fold structures of rock salt deposits and of Archaean formations in many parts of the world, can be explained only by assuming that these rocks flowed under small stress differences like viscous fluids. The Archaean formations become highly mobile when buried deep within the crust, and

[1] See e.g. Dickinson, S. B., 'The Structural Control of Ore Deposition in some South Australian Copperfields': *Bull. No. 21, Geol. Surv. South Australia.*

[2] Aubert ('Le Jura': *Geol. Rundsch.*, Vol. 37, 1949, pp. 2–17) has stressed the importance of highly mobile beds as active transmitters of stress in the Juras, and the drag effects in salt domes indicate that mobile rocks cannot be regarded as playing only a passive rôle in tectonics.

also in other ways subjected to conditions favouring recrystalization and crystal-plasticity, but salt deposits are highly plastic even in the upper crustal levels. As is shown by its behaviour in the Alsatian and Hallstatt salt mines,[1] rock salt will flow when buried beneath 2,000 feet or less of sediments. It has been demonstrated that the elastic limit of crystalline sodium chloride is lowered with increase of temperature, and also on immersion in water, when its ductility is, in addition, increased.[2] The readiness with which rock salt will flow is well illustrated by the remarkable salt glaciers of Laristan,[3] and the

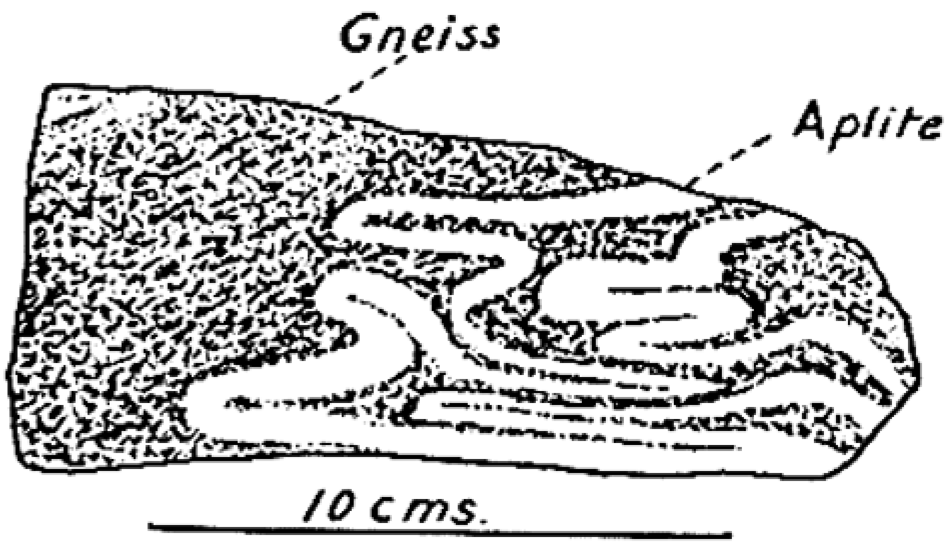


FIG. 56.—FLOW FOLDING (PTYGMATIC FOLDING) SHOWN BY AN APLITE VEIN IN PRE-CAMBRIAN GNEISS

(After Sederholm, 'Über Ptygmatische Faltung': *Neues Jahrb.*, B.B. 36, 1913, pp. 491–512)

structures of salt domes, which have been very fully investigated because of their importance in oil-field geology, also throw light on this point. Beds containing gypsum or anhydrite are also highly plastic under similar conditions.[4] Many folds in slumped rocks are also flow folds.

[1] Lees, G. M., 'Some Depositional and Deformational Problems': *Journ. Inst. Petrol. Technol. Lond.*, Vol. 17, 1931, p. 273.

[2] Ewald, W., and M. Polanyi, *Zeitschr. f. Physik*, Bd. 28, 1924, pp. 29–50. Houwink, R., *Elasticity, Plasticity and Structure of Matter*: Cambridge, 1937, pp. 101–8.

[3] Harrison, J. V., 'The Geology of Some Salt-Plugs in Laristan (Southern Persia)': *Quart. Journ. Geol. Soc.*, Vol. 86, 1930, pp. 463–552. Wade (*Journ. Inst. Petrol. Technol. Lond.*, Vol. 17, 1931, p. 357) has also described gypsum glaciers, occurring on the northern shores of the Red Sea.

[4] See Griggs, D., 'Experimental Flow of Rocks under Conditions favouring Recrystallization': *Bull. Geol. Soc. Amer.*, Vol. 51, 1940, pp. 1001–34, for gypsum, and remarks on the Juras (pp. 58–9) for anhydrite.

Compound Folds.—In zones of folded rocks, inhomogeneity of stress and rock matter result in the possibility that individual folds may arise by one or other or by a combination of the above mechanisms. An important type involving flexing of competent strata in buckle-folds and slipping of interbedded incompetent strata along planes of false cleavage is the *flexural-slip* type of fold[1] (Fig. 67). Flow folds may be associated with buckle-folds, as in the Juras, or with bending folds as in the doming of strata over a salt plug.

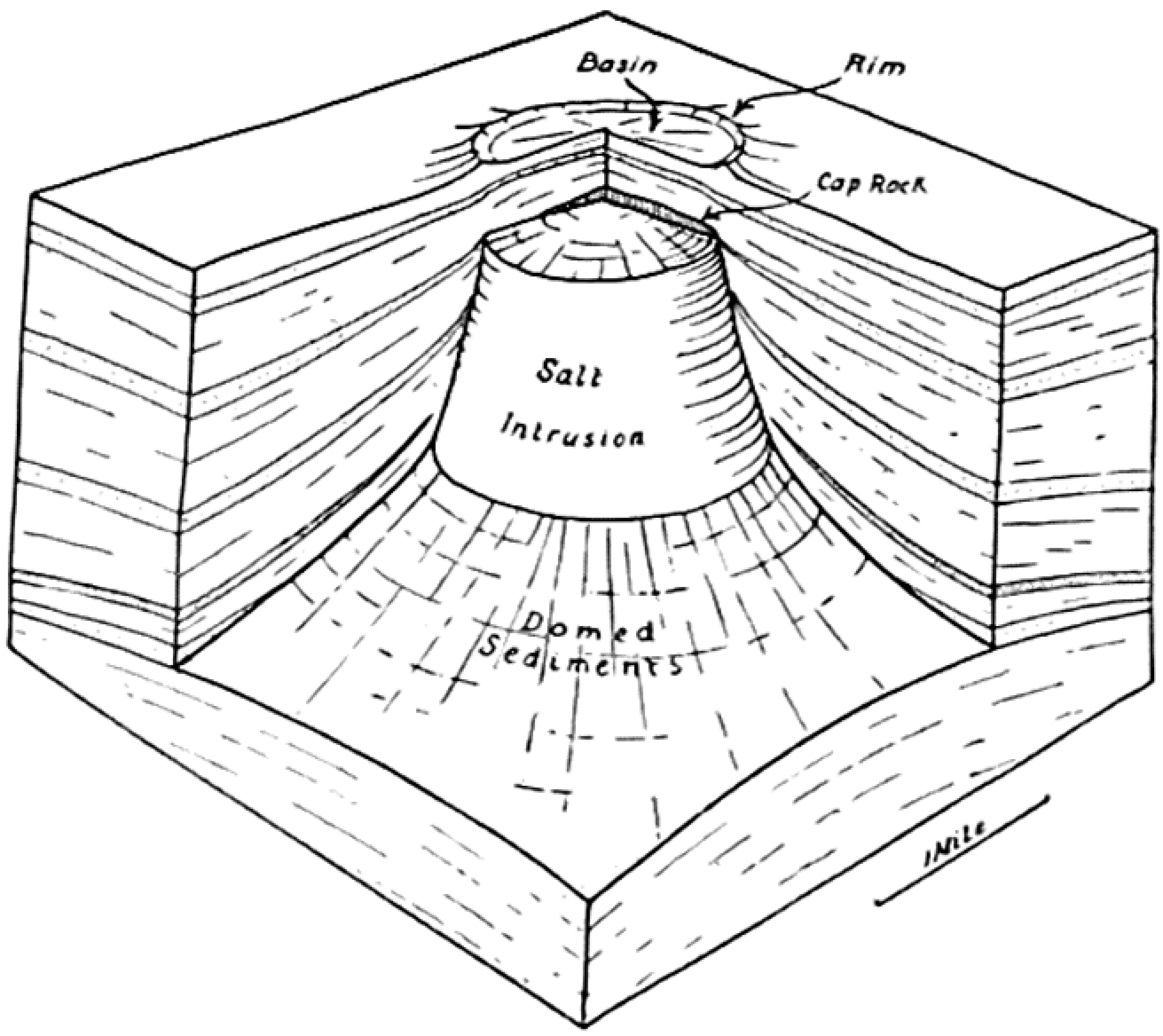


FIG. 57.—DISSECTED BLOCK DIAGRAM OF A SALT DOME, SHOWING THE DOMING OF THE SEDIMENTS AROUND THE SALT INTRUSION, THE LIMESTONE CAP ROCK, AND THE TOPOGRAPHIC BASIN AND ANNULAR RIDGE PRODUCED BY EROSION OF THE OVERLYING BEDS AT THE SURFACE

(Simplified after Carlton in *Structure of Typical American Oilfields*, Vol. II, 1929)

Salt Domes.

In many regions where salt deposits occur, the salt, although it must originally have possessed the sheetlike form of

[1] Knopf, E. F., 'Petrotectonics': *Amer. Journ. Sci.*, Ser. 5, Vol. 25, 1933 pp. 433–70 (see pp. 464–7); 'Structural Petrology': *Mem. Geol. Soc. Amer.*, No. 6, 1938, *passim.*

normal sedimentary strata, is found to have accumulated into masses. These may be circular, elliptical, elongated, or irregular in plan, and typically they penetrate the beds that would normally overlie the salt. Buried or partially uncovered ridges of salt, of which one axis in plan is considerably longer than the other, are termed *salt anticlines*, while *salt plugs* or *stocks* are subcircular in plan. The general term *salt dome* is used to include these and the less-regular types of salt bodies. Normally, the overlying beds are domed up over the salt masses, the wall rocks are strongly dragged upwards (Fig. 57), and are also

FIG. 58.—SECTION ACROSS A SALT DOME NEAR HANOVER
(After Seidl)

The basement rocks are not folded, but the salt shows complex folding and thrusting, and has domed up the overlying beds.

often brecciated, while the salt exhibits complex folding and minor shearing, excellently shown in the salt mines of Saxony (Fig. 58).[1] At the summit of many salt domes is a *cap rock*, consisting of calcite, gypsum, and anhydrite, often with some sulphur.

It is now firmly established that salt domes are intrusive bodies, which have actively penetrated the sedimentary strata

[1] Important works on salt tectonics are—*The Geology of Salt Dome Oil Fields*: published by Amer. Assoc. Petrol. Geologists, 1926. 'Symposium on Salt Domes': *Journ. Inst. Petrol. Technol. Lond.*, Vol. 17, 1931, pp. 252–383. *Problems of Petroleum Geology*: published by Amer. Assoc. Petrol. Geologists, 1934. Fulda, E., 'Salztektonik': *Zeit. d. deutsch. geol. Gesellsch.*, Abt. 79, 1927, pp. 178–96. Also various articles in *The Science of Petroleum*: Oxford, 1938.

that formerly overlay the original salt deposit.[1] By Stille it is held that the rise of the salt is mainly due to the effects of lateral compression in the crust, the intrusive salt masses being regarded as extreme examples of diapiric folds which have been aided in their formation by the mobility of the salt. This concept is doubtless applicable in regions of compression such as the Roumanian oil-fields, but many salt domes are unrelated to tectonic folds, and the evolution of these is pictured as follows. Irregularity in the pressure exerted on the salt deposit by the overlying strata results first in lateral flow of the salt, which accumulates as ridges beneath places where the pressure is low. Such irregularities may arise from variation in the thickness of the salt and of the overlying sediments, from structural weaknesses in the overburden, from erosion, or from arching of the salt or the overlying beds. Once the salt has accumulated into masses, hydrostatic forces, due to the low specific gravity of the salt compared with that of the normal sediments above it, cause the masses to rise through the overlying sediments, as oil rises through water. On this hypothesis, the salt will cease to rise when the hydrostatic forces are no longer sufficient to overcome the resistance offered, or when the supply of salt is exhausted. The possibility is thus indicated that salt masses without roots may exist. Experimental studies lend support to the suggestion that it is the low specific gravity of the salt deposits which is mainly concerned with the formation of intrusive salt domes,[2] although both the salt and the surrounding rocks that are shouldered aside, must be plastic for intrusion to take place. The occurrence of these structures in regions where the rocks are very little disturbed, as in the Gulf Coast region of North America, also points to the probable validity of this theory, though in the Roumanian oil-fields, where the tectonics are complex, lateral pressure may actually dominate in the control of salt structures.

[1] Stille, H., *Geology of Salt Dome Oil Fields*: 1926, p. 142.

[2] Escher, B. G., and P. H. Keunen, 'Experiments in Connexion with Salt Domes': *Leid. Geol. Med.*, Deel 3, 1929, pp. 151–82. Link, T. A., 'Experiments Relating to Salt Dome Structures': *Bull. Amer. Assoc. Petrol. Geologists*, Vol. 14, 1930, p. 4. Nettleton, L. L., *ibid.*, Vol. 18, 1934, p. 1175.

3. *MINOR STRUCTURES IN FOLDED ROCKS*

Drag Folds.—As has already been noted (p. 81), the development of parallel folds in competent strata is accompanied by slip of the beds relatively to each other, along the bedding planes. Evidence of this *bedding plane slip* is often afforded by slickensides. Where competent and incompetent strata are interbedded, the movements of the former relatively to each other subject the incompetent strata to shearing stress, which may cause small folds, called *drag folds*, to develop in them[1] (see Fig. 59). The apices of drag folds lie at right-angles to the direction of the differential slip between the competent beds, so that in the case of buckle-folds, in which the upper of two adjacent competent beds moves over the lower towards the crest of an anticline, the trend of the axial lines of drag folds should be parallel to that of the fold on whose limbs they occur. The direction and angle of pitch of the drag folds and the major fold should in such cases correspond.

It has been stated that 'the degree and direction of the pitch of a fold are often indicated by those of the axes of the minor plications on its sides' (Pumpelly's rule).[2] This rule, as originally formulated, does not refer to drag folds in the sense defined above, but to the minor folds of an anticlinorial or synclinorial structure. It was formulated by Dale under Pumpelly's guidance, and was stated to be 'applicable primarily to the study of the metamorphic rocks of Mount Greylock, and then to a large part of the Taconic region and to similar rocks and regions'. Later workers have extended the rule to include drag folds as well as minor folds,[3] in the belief that minor folds are caused by drag between major competent formations in anticlinoria and synclinoria.[4] The universal applicability of this

[1] van Hise, C. R., and C. K. Leith, 'The Geology of the Lake Superior Region': *U.S. Geol. Surv.*, Mon. No. 52, 1911, p. 123.

[2] Pumpelly, R., J. E. Wolff, and T. N. Dale, 'Geology of the Green Mountains in Massachusetts': *U.S. Geol. Surv.*, Mon. No. 23, 1894, p. 158.

[3] Leith, C. K., *Structural Geology*: New York, 1923, pp. 176–81. Nevin, C. M., *Principles of Structural Geology*: New York, 1931, p. 74. Willis, B., and R. Willis, *Geologic Structures*: New York, 1934, pp. 97–9.

[4] van Hise, C. R., and C. K. Leith, 'The Geology of the Lake Superior Region': *U.S. Geol. Surv.*, Mon. No. 52, 1911, p. 253.

latter concept lacks confirmation, but nevertheless Pumpelly's rule does apply to the majority of minor folds in many regions of complex structure. Exceptions are, however, so numerous that the rule must be used with discretion as a guide to major structures. Furthermore, the practical value of extending it to include true drag folds is doubtful. The attitude of these depends entirely upon the direction of relative movement of the thick competent strata, and in regions whose tectonic history has been complex, such movements will certainly have taken place in several directions. van Hise and Leith state that in the Vermillion district the axes of drag folds lie 'in any direction in the plane of bedding', and Derry[1] has described steep-

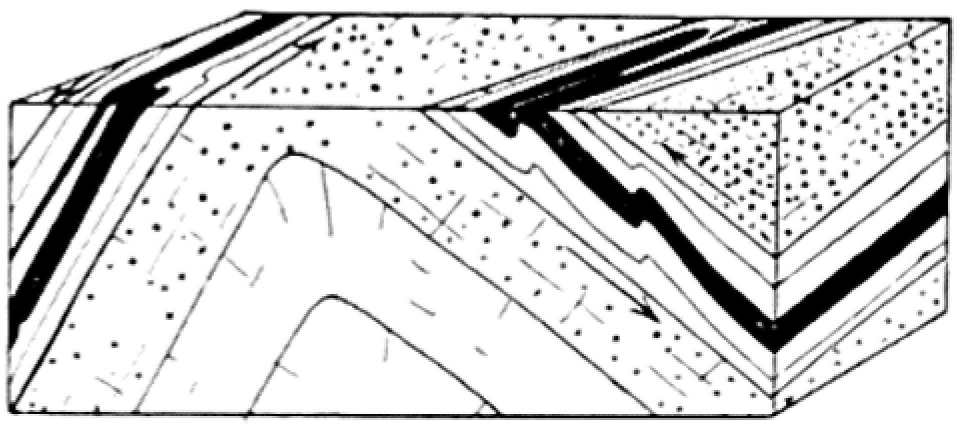


FIG. 59.—DRAG FOLDING IN FOLDED STRATA

On the right are shown congruous drag folds caused by bedding plane slip in the direction of the arrows: on the left, incongruous drag folds due to horizontal bedding plane slip.

pitching drag folds, independent of the major folding, as well as drag fold dependent on that folding, in the Pre-Cambrian rocks of the Canadian Shield. In the Lower Palaeozoic rocks of Victoria, similar steep-pitching drag folds occur, related to nearly horizontal bedding plane slip developed in already steeply dipping beds. It may be useful to apply the term *congruous* to those minor folds and drag folds that agree with Pumpelly's rule, being presumably dependent on the major folding forces, and to refer to those that do not, whether dependent or independent, as *incongruous* (see Fig. 59).

On the assumption that an observed drag fold is congruous

[1] Derry, D. R., 'Some Examples of Detailed Structure in Early Pre-Cambrian Rocks of Canada': *Quart. Journ. Geol. Soc.*, Vol. 95, 1939, pp. 109–34.

it may be used in the field to indicate the direction in which the neighbouring synclinal and anticlinal axes lie. The acute angle between the axial plane of a drag fold and the bedding planes of an adjacent competent bed points in the direction in which the competent bed has moved, and, remembering that the upper of any two competent beds moves towards the crest of an anticline and the lower towards the trough of a syncline during folding, the position of the anticline and syncline can be deduced, as in Fig. 60.

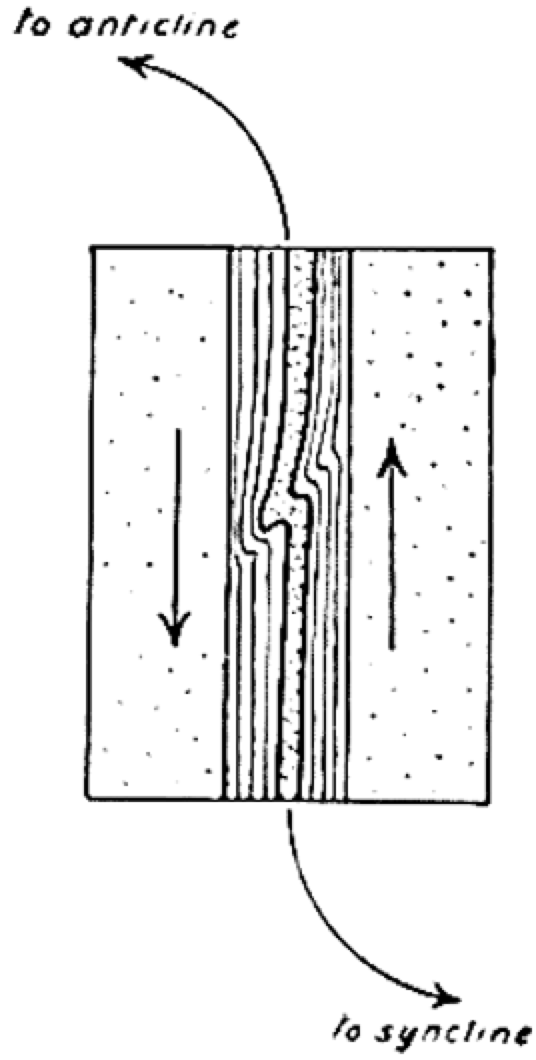


FIG. 60.—DRAG FOLDS (ASSUMED TO BE CONGRUENT) IN INCOMPETENT BEDS LYING BETWEEN COMPETENT SANDSTONES

The direction of bedding plane slip, inferred from the attitude of the axial plane of the drag fold, is in the direction of the arrows, so that the sandstone on the right is the youngest bed.

Joints.—A joint is a fracture in a rock mass, along which there has been extremely little or no displacement. At the surface, joints often become open fissures as a result of weathering, but below the zone of weathering they are closed, and sometimes sets of joints in the surface rocks are not represented in depth. Joints develop in sedimentary rocks from a variety of causes, including shrinkage accompanying dehydration, expansion due to weathering, tectonic deformation, and tidal effects in the crust. The jointing of igneous rocks is dealt with in Chapter VI. Joints are said to form *sets* when they are arranged in one series which is parallel in dip and strike over a considerable area, or *systems* when there are two or more sets intersecting at a more or less constant angle. The analogy of joint systems with the shear planes and tension gashes of experimentally deformed solids has long been recognized,[1] and

[1] Daubrée, A., *Études Synthétiques de Géologie Expérimentale*: Paris, 1879; 'Application de la Methode expérimentale a l'étude des Déformations et des Cassures terrestres': *Bull. Soc. Géol. France*, Vol. 7, pp. 108–41. Daubrée's

tectonic joints may be classified as *shear* (or *slip*) *joints* and *tension joints*, according to their nature.[1]

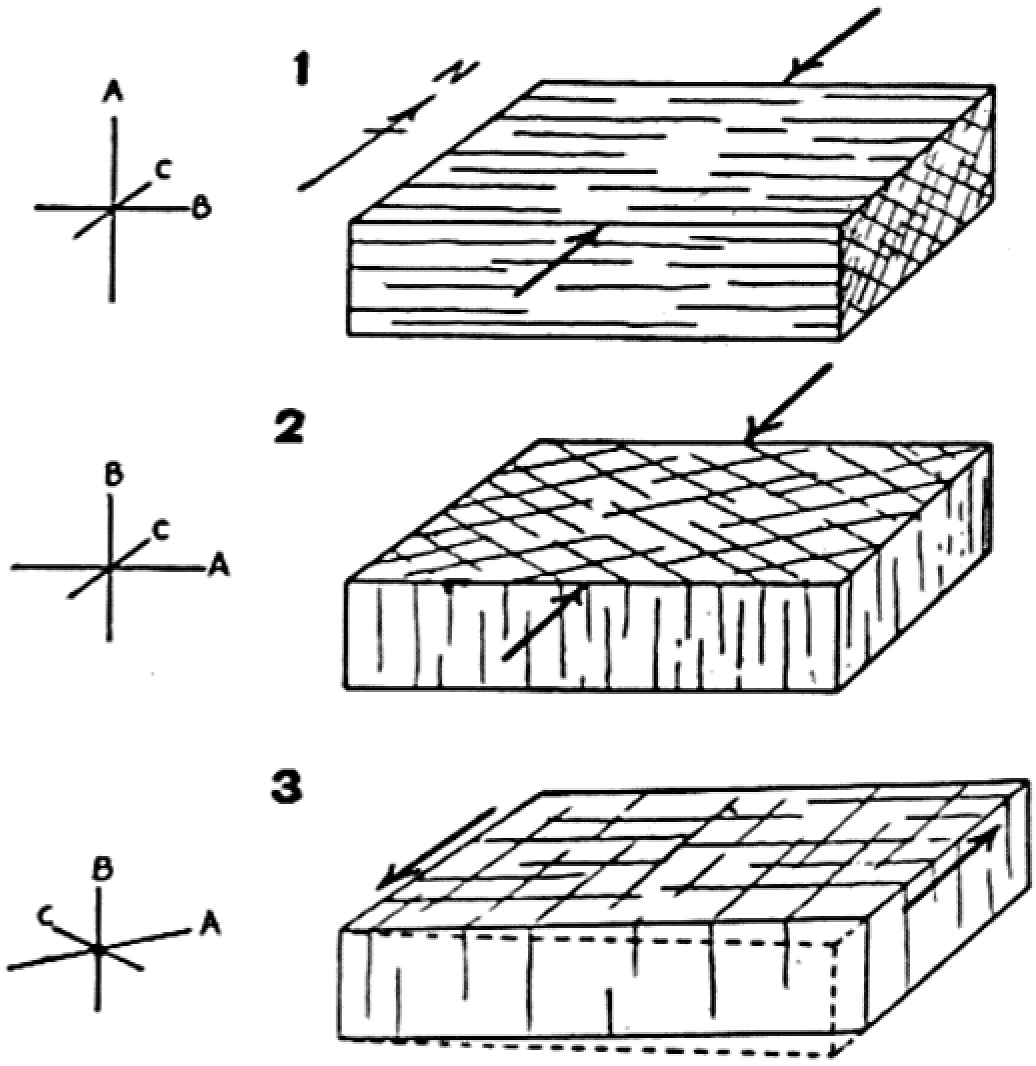


FIG. 61.—SYSTEMS OF SHEARING JOINTS

(Adapted from Nevin, *Principles of Structural Geology*)

1. Joints are caused by horizontal compression (N.–S.), with the direction of easiest relief upwards. A, B, C, are the principal axes of the strain ellipsoid.
2. Joints are caused by horizontal compression (N.–S.), with the direction of easiest relief horizontal (E.–W.).
3. Joints are caused by horizontal shearing stress, with the direction of easiest relief horizontal.

Shear joints are either shearing planes along which the differential slip has been of microscopical amount, or potential shearing planes which have become visible as a result of

work should be used with caution in the light of modern knowledge—see (p. 42), also Bucher, W. H., 'The Mechanical Interpretation of Joints': *Journ. Geol.*, Vol. 28, 1920, pp. 707–30; *ibid.*, Vol. 19, 1921, pp. 1–28.

[1] Cloos, H., *Einführung in die Geologie*, pp. 224–30. Nevin, C. M., *Principles of Structural Geology*, pp. 138–54. Leith, C. K., *Structural Geology*, pp. 29–64. Willis, B., and R. Willis, *Geologic Structures*, pp. 114–40. Bucher, W. H., *op. cit.* 1920. Sheldon, P., 'Some Observations and Experiments on Joint Planes': *Journ. Geol.*, 20, 53–183, 1912. Swanson, C. O., 'Notes on Stress, Strain, and Joints': *ibid.*, Vol. 35, 1927, pp. 193–223.

readjustments taking place in the deformed rocks as a result of weathering, or of fatigue under alternating stress.[1]

In field interpretations of joint sets and systems it is important to remember that the direction of the axis of mean strain is given by the intersection of complementary sets of shearing planes, and that the tension joints or gashes form at right angles to the axis of greatest strain, in the plane of the axes of least and mean strain. Interpretations of systems of major joints on these principles are shown in Fig. 61. In the first case, pressure acts from north to south on rocks in which the easiest relief is in a vertical direction. In the second, the direction of easiest relief is assumed to be in an east–west direction, vertical joints resulting under compression from north and south. Finally, shear in a horizontal plane, with easiest relief also in this plane, will result in a similar joint system, and it is therefore not possible to infer the nature of the forces which have acted on the rocks from a consideration of jointing alone, though taken in conjunction with all the other evidence, jointing may be useful in this regard.[2]

Jointing in Closely Folded Sediments.—The jointing is usually very complex in closely folded sediments. B. Willis and R. Willis[3] have attempted to differentiate between the type of jointing in folds caused by direct compression on the one hand, and by horizontal shearing stress on the other. The theoretical interpretation of joints in folded rocks is, however, not yet based on sufficiently sound principles to allow deductions to be made from joints concerning the origin of the folds. A more empirical treatment therefore appears preferable.

Apart from the regional joints, treated above, there are local joints obviously related geometrically to individual folds. At

[1] For a discussion of the latter see Kendall, P. F., and H. Briggs, 'The Formation of Rock Joints and the Cleat of Coal': *Proc. Roy. Soc. Edinburgh*, Vol. 53, 1933, pp. 164–87.

[2] Wager, L. R., 'Jointing in the Great Scar Limestone of Craven, and its Relation to the Tectonics of the Area': *Quart. Journ. Geol. Soc.*, Vol. 87, 1931, pp. 392–424. See also Bucher, W. H., *Journ. Geol.*, Vol. 28, 1920, pp. 707–30; and Kendall and Briggs, *Proc. Roy. Soc. Edinburgh*, Vol. 53, 1933, pp. 164–87, for the effects of torsion.

[3] *Geologic Structures*: New York, 1934, pp. 94–105.

crests and troughs, for instance, competent strata often exhibit radially arranged joints, generally regarded as being tension joints caused by the stretching of the outer sides of the beds in

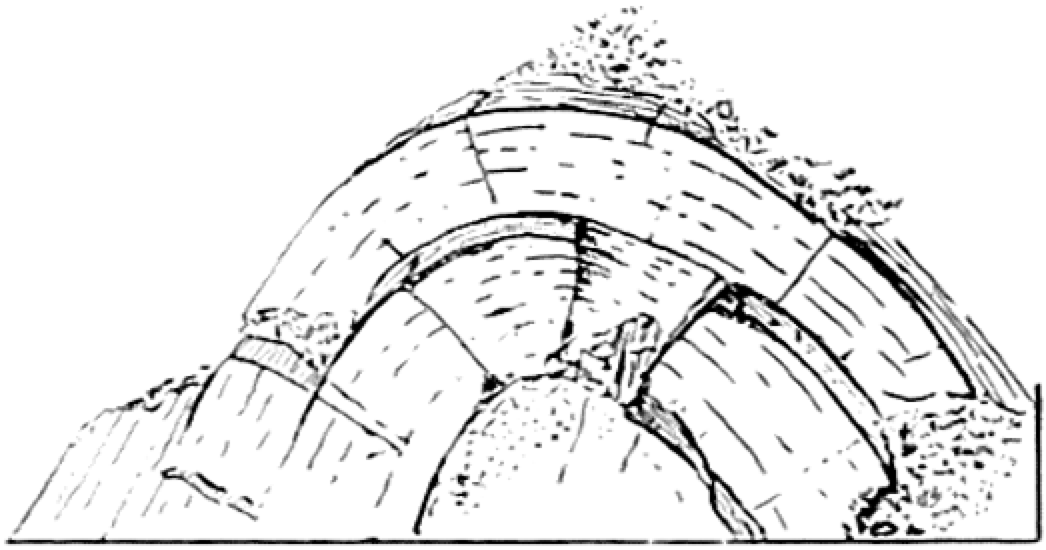


FIG. 62.—RADIAL TENSION JOINTS IN SANDSTONE AT AN ANTICLINAL AXIS. CASTLEMAINE, VICTORIA

(From a photograph in *Mem. Geol. Surv. Vict.*, No. 2, 1903: The Castlemaine Goldfield)

the flexures (see Fig. 62, and p. 41). These are strike joints. Other strike joints, not radially arranged but forming parallel sets, represent incipient shearing planes, probably induced by

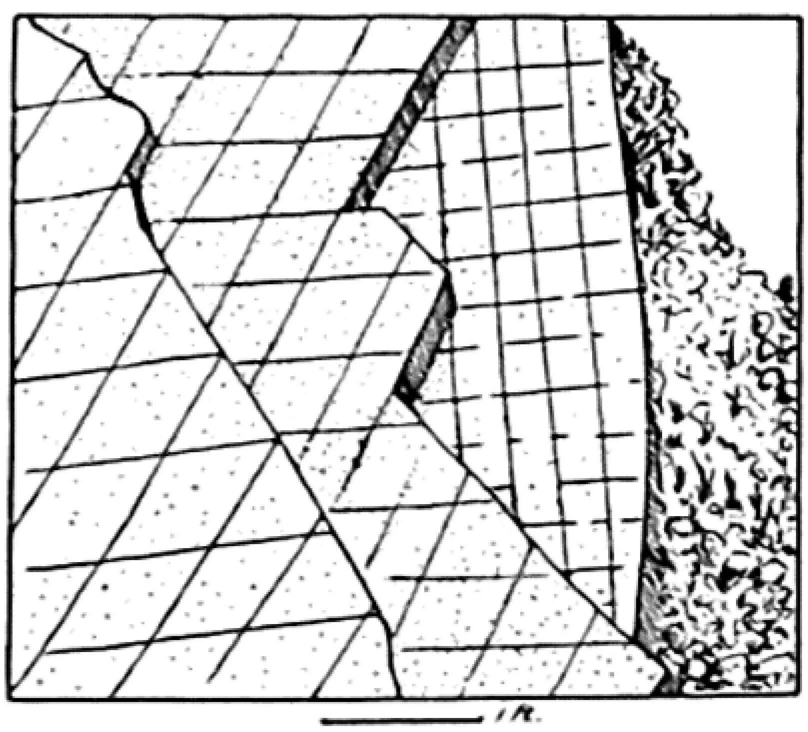


FIG. 63.—INTERSECTING JOINT SETS IN THREE SANDSTONE BEDS AT STUDLEY PARK, VICTORIA

The joints in each bed are independent of those in the others.

the regional compression at right angles to the axes of the folds. Closely folded beds are also often divided into parallelepipeds by the parting planes afforded by two intersecting sets of joints

and the stratification planes[1] (Fig. 63). The pattern of these joints strongly suggests that they are incipient shearing planes: in many instances they intersect along lines at right angles to the stratification planes, and the angle between them is different in each bed. Willis and Willis[2] classify them as compression joints, but their origin is more probably connected with weathering. Other examples of intersecting shear joints can be related to faults transecting the folds, and thus of later origin than them. In these, the intersection of the shearing planes may be oblique

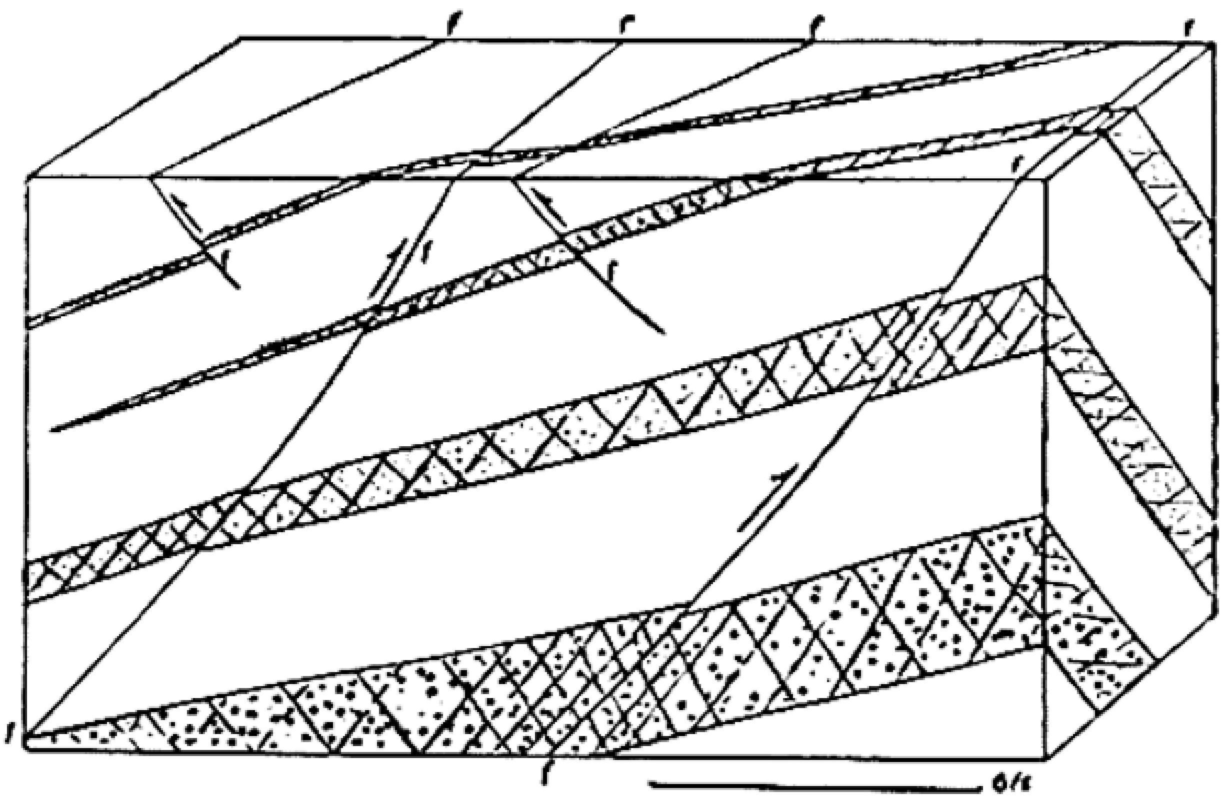


FIG. 64.—SHEAR JOINTS COGNATE WITH SHEAR THRUSTS (*f-f*) IN SILURIAN SANDSTONES AT STUDLEY PARK, VICTORIA. INTERBEDDED MUDSTONES ARE NOT JOINTED

to the bedding, and the joints may be either parallel to the faults (Fig. 64), being probably caused by the same stresses, or at an angle to them (see pp. 131–5), and due to locally developed stresses connected with the faulting. Joints that lie approximately at right angles to the crests and troughs of folds and cut across several beds may be termed *transverse joints*, (Fig. 65). They are probably tension joints caused by the stretching of crests and troughs that are arched longitudinally. In the field transverse joints are useful since they indicate the pitch of folds, to which they are normal.

[1] A third subsidiary joint set is also often present.
[2] *Geologic Structures*: New York, 1934, pp. 114–16.

The possible complexity of minor structures accompanying folding is well illustrated in experiments with clays.[1] A cake of clay, subjected to a direct push from one side, is thrown into folds striking at right angles to the active pressure, but these folds are transected by complementary shearing planes of the nature of normal faults, trending at right angles to the folds (see Plate IV). The axial lines are offset in places by vertical shearing planes making an angle of about 50° with them. These directions of shearing may be developed either as faults or joints. It may be remarked that there are two reasons why more than one set of shearing planes arises in an experiment carried out with homogeneous material. The first is that the

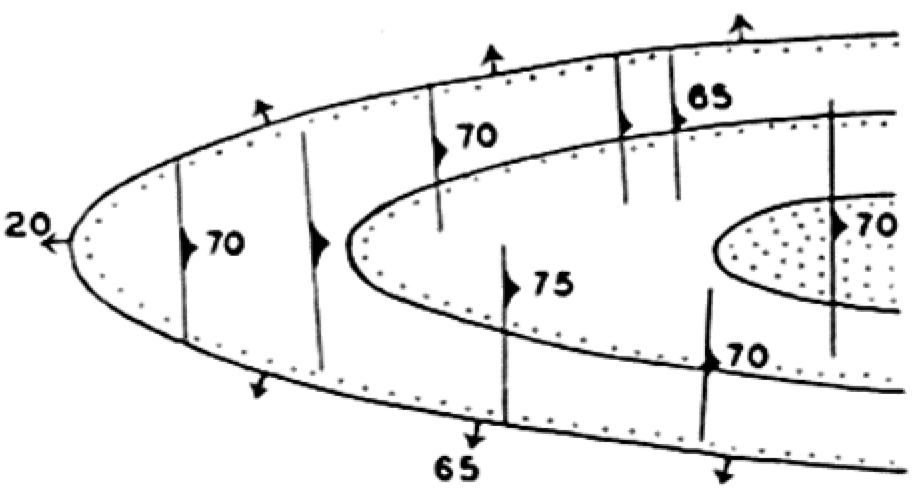


FIG. 65.—MAP OF THE 'NOSE' OF A PITCHING ANTICLINE, SHOWING TRANSVERSE JOINTS LYING APPROXIMATELY AT RIGHT ANGLES TO THE CREST OF THE FOLD

strain is heterogeneous, especially after some of the clay has piled up above the original surface of the cake. The second is that folded portions bounded below by strongly developed shearing planes are no longer subjected to the full effects of the compression, and subsidiary shearing planes due to the arching of the fold crests can thus be developed.

Cleavage.—Rocks that possess a cleavage may be split into thin sheets along parallel or subparallel planes that are of secondary origin and are formed as a consequence of metamorphism. Cleavage is independent of original structures such as bedding, and it generally maintains a certain constancy of dip and strike irrespective of the folding of the strata, so that

[1] Cloos, H., 'Zur Tektonischen Stellung des Saargebietes': *Zeits. d. Deutsch. Geol. Ges.*, Vol. 85, 1933, pp. 307–15; *Einführung in die Geologie*: Berlin, 1936, pp. 272–9.

it may intersect the bedding at any angle. Cleavage is determined either by mechanically formed planes of weakness such as shear, flow or knick-planes, or by the parallel orientation of flaky mineral particles formed chiefly during recrystallization, and in most, if not all rocks possessing good cleavage, both factors have operated.

Shearing may take place along one set of surfaces only, in which case it lies at an angle to the A–B plane of the strain ellipsoid, appropriate to the rock under the conditions of deformation that obtain. Again, shearing may occur along two conjugate directions, in which case the actual displacement of any portion of the rock is determined by the sum total of movements on both, and lies in some position intermediate between the conjugate shear surfaces. Cleavage may follow one dominant shear direction or may correspond with the direction of elongation ('flow') in a rock, or with other structures such as knick-planes; thus its attitude in relation to the strain ellipsoid is variable. Evidence for the translations involved may be obtained from the displacement of planes such as bedding, foliation, or earlier cleavage; from the deformation of fossils, oolite grains, pebbles and the like, from consideration of the geometry of parallel-sided strata, and from direct observation of the shearing and fracturing of mineral grains (Fig. 107).[1]

Interpretation of observed deformations in terms of the strain ellipsoid is difficult for many reasons. Firstly, a given strain may arise in various ways (see p. 27); secondly, the initial attitude of structures may be changed by later movements; and again, the physical properties of rocks at the stage of cleavage development are unknown. For instance, it is thought that planes which, from the displacement of bedding or foliation along them, appear to be simple shearing planes (Fig. 55) may in fact be analogous to flow planes such as are

[1] In addition to references on page 105, see Harker, A., 'On Slaty Cleavage and allied Rock Structures': *Rept. Brit. Assoc. Adv. Sci.*, 1885 (1886), p. 813; Hills, E. S., and D. E. Thomas, 'Deformation of Graptolites and Sandstones in Slates from Victoria, Australia': *Geol. Mag.*, Vol. 81, 1944, pp. 216–22. Fellows, R. E., 'Recrystallization and Flowage in Appalachian Quartzite': *Bull. Geol. Soc. Amer.*, Vol. 54, 1943, pp. 1400–32.

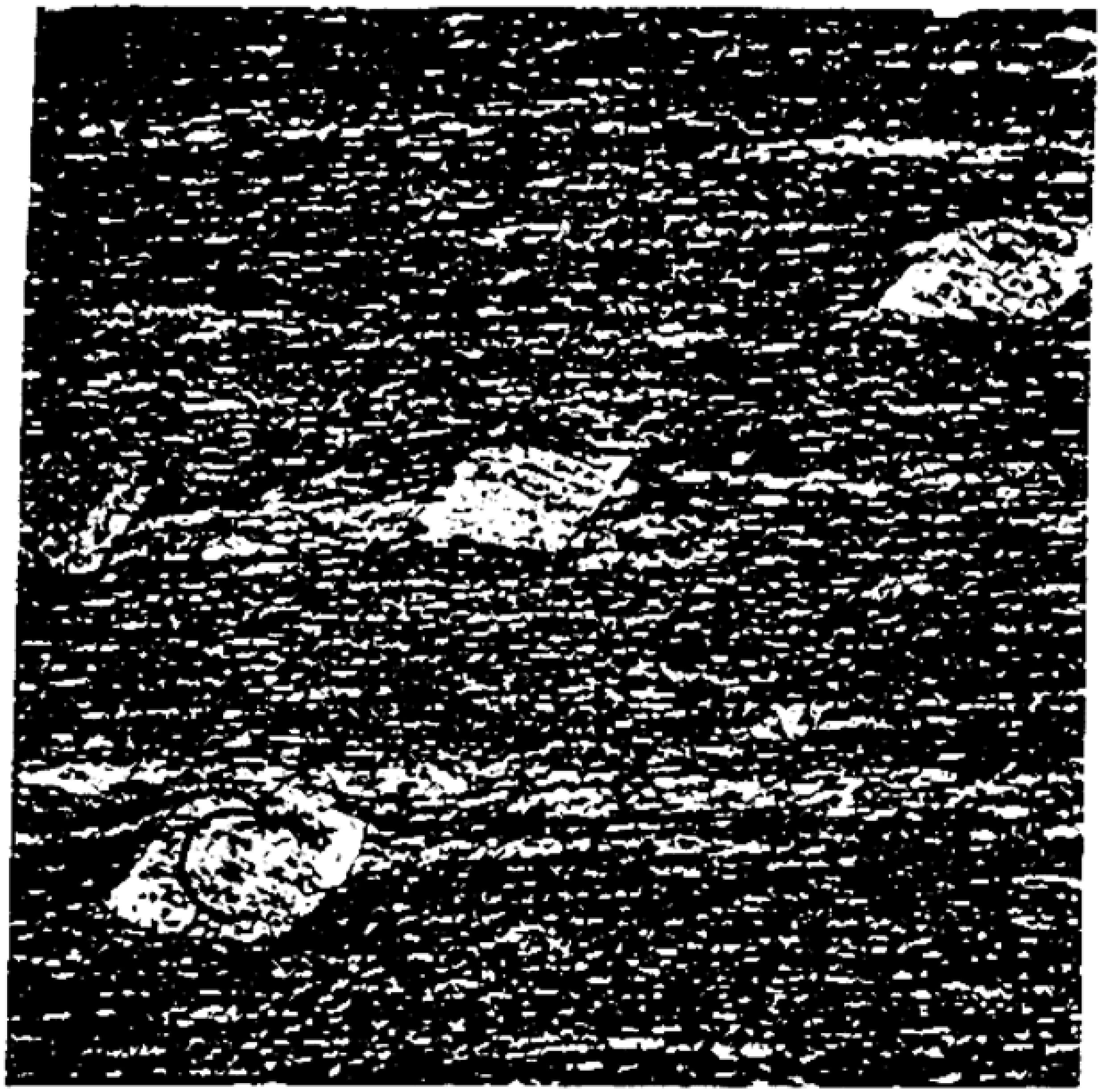
formed in rolled or extruded metals, in wet clay, or in liquids, and thus may lie normal to the axis of compression.[1]

The orientation of new minerals formed during chemical reconstitution is controlled by the stress conditions, and by directional influences such as bedding planes, shearing, and flow surfaces in the rock, so that again the resultant cleavages have various angular relationships with the strain axes (see pp. 159–61). However, since cleavage is by definition independent of bedding, examples of recrystallization, following original bedding as in some schists, are not properly referable to cleavage but may be termed *bedding-schistosity*, or, at a lower metamorphic grade, *bedding-fissility*.

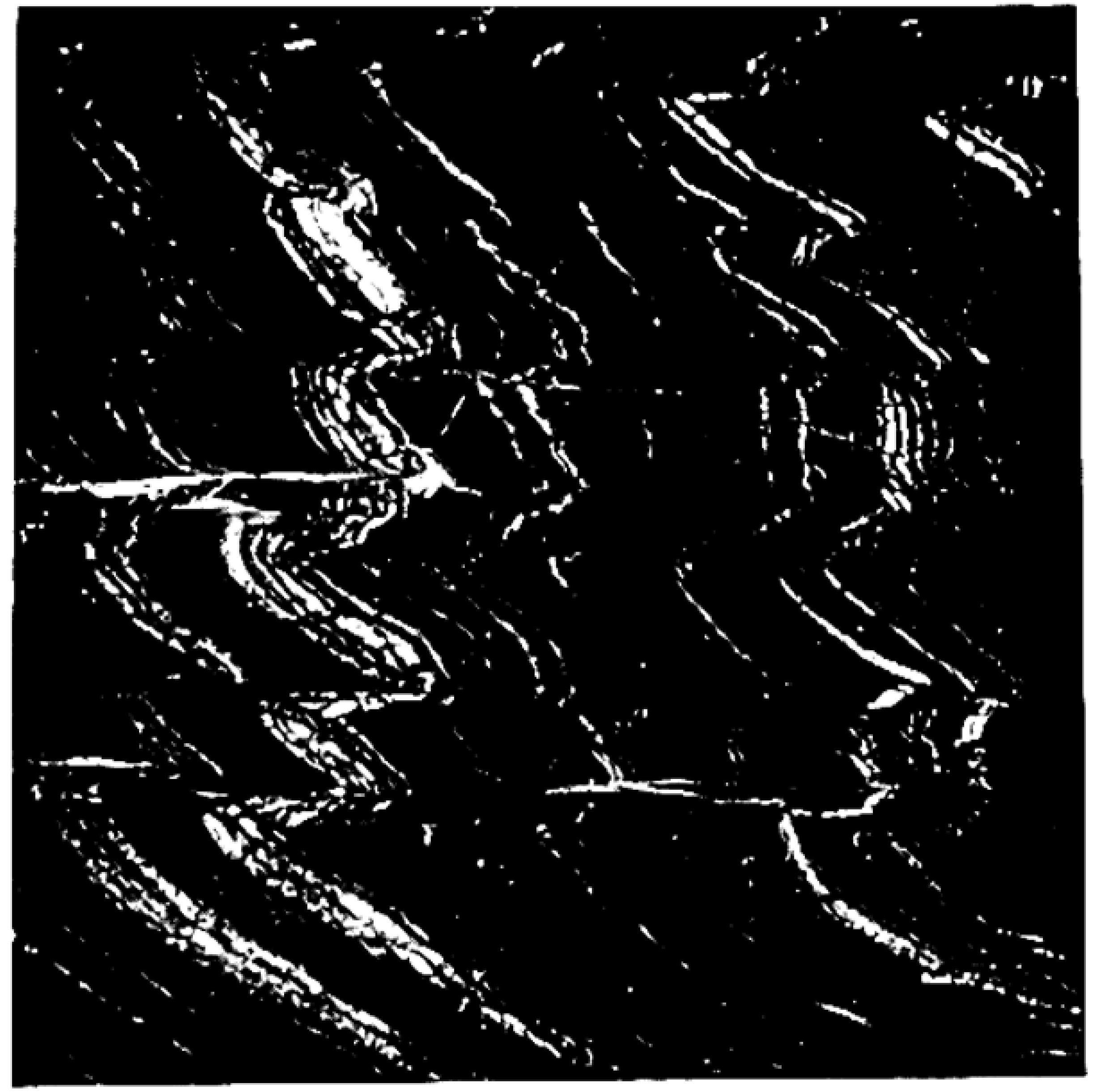
Although it is clear that 'cleavage' includes structures formed in a variety of different ways, it bears a close relationship in all to the movements of the parts of the rock concerned, and similar rock-types in comparable tectonic settings typically develop cleavage in somewhat similar ways, so that it is of great importance in mapping.

Slates, for which the term cleavage was originally coined, are formed chiefly from argillaceous sediments or fine-grained tuffs. Reconstitution affects the original clay minerals, the hydrolysates and carbonaceous constituents, but quartz and much of the original mica are in general mechanically deformed rather than recrystallized. The specific gravity of slate is higher than that of the parent rock, partly because of reduction in the percentage of voids and partly because of loss of combined water during recrystallization. The majority of secondary lamellar minerals in slates (e.g. mica, chlorite, hematite and graphite) lie with their plane surfaces and good cleavages in parallel orientation, due either to their growth in some definite space relationship to the strain or to the stress-axes, or as a result of rotation into parallelism by drag along a dominant set of shearing or flow planes in the plastic rock mass. Fig. A, Pl. III, illustrates the microscopical structure of

[1] White, W. S., 'Cleavage in East-Central Vermont': *Trans. Amer. Geophys. Union*, Vol. 30, 1949, pp. 587–94. Cloos, E., 'Oolite Deformation in the South Mountain Fold, Maryland': *Bull. Geol. Soc. Amer.*, Vol. 58, 1947, pp. 843–918.



A. CLEAVAGE IN SLATE, CASTLEMAINE, VICTORIA. × 30



B. CLEAVAGE IN LAMINATED SLATE, DONNELLY'S CREEK, VICTORIA. ½

slate in which the rotation of the calcite porphyroblasts, revealed by the lines of inclusions which were originally parallel with the cleavage, shows that differential slip has taken place along this direction during the growth of the calcite crystals. Such contemporaneous recrystallization and deformation is referred to as syntectonic crystallization, or paracrystalline deformation (see pp. 167–8).

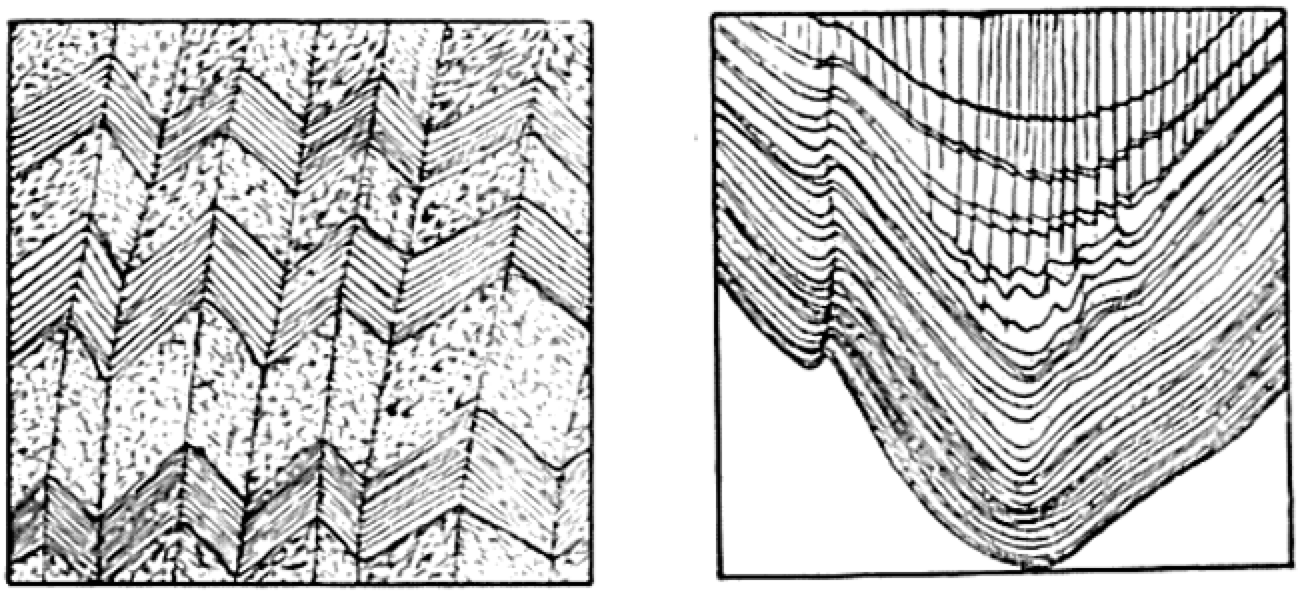
Mineralogical reconstitution is effective in all slates; potential voids or ruptures are filled as they form with secondary quartz and chlorite, as is well shown in the 'eyes' or 'pressure fringes' around hard crystals such as pyrites (see p. 169), but mechanical effects are also important. Individual planes of shearing or flowage are generally discernible under the microscope, although they may be extremely closely spaced, and again, the shearing or flexing of sandy or coloured stratification-layers and laminae along the cleavage is characteristic.[1] In beds graded from an arenaceous base to an argillaceous top, cleavage is curved in the manner shown in Fig. 67, A, deviating sharply in passing upwards from an argillaceous top to an adjacent arenaceous base. Cleavage in sharply bounded arenaceous strata interbedded in slate deviates abruptly from that in the slate except where cleavage and bedding are at right angles to each other, as they normally are at fold axes. The angular relationships of cleavage are therefore connected with the mineral composition of the rocks and their situation in relation to the geometry of folds, which is more fully discussed later.

Types of Cleavage.—A variety of types of cleavage has been recognized by different authors, but two contrasted types, flow and fracture cleavage respectively, are classic concepts in North American structural geology.[2] In *flow cleavage* the rock is thoroughly reconstituted and its cleavability depends on

[1] Excellent illustrations in rocks possessing typical slaty cleavage are given by Dale, T. N., 'The Slate Belt of Eastern New York and Central Vermont': *19th Ann. Rept. U.S. Geol. Surv.*, Pt. 3, pp 153–307, and such phenomena are common in all slate belts.

[2] Leith, C. K., 'Rock Cleavage': *U.S. Geol. Surv.*, Bull. 239. An excellent account of cleavage is given by Wilson, G., 'The Relationship of Slaty Cleavage and Kindred Structures to Tectonics': *Proc. Geol. Assoc.*, Vol. 57, 1946, pp. 263–302 (with Bibliography).

the parallel orientation of lamellar minerals of micaceous habit. It will split along any arbitrarily chosen plane parallel with the direction of mineral orientation, and the term is therefore particularly applicable to transverse schistosity, since in schists individually distinct planes of shearing or flowage are obliterated in favour of an overall directional texture that extends even to the crystal lattices of component minerals. As explained above, its use for bedding schistosity is not advocated herein. Ancillary notions that have become attached to the flow cleavage concept by various authors are that it forms in the A–B



A B

FIG. 66.—A. CLEAVAGE PARALLEL TO THE AXIAL PLANES OF PUCKERS IN MICA SCHIST. CASTERTON, VICTORIA. (× 1 APPROX.)
B. CLEAVAGE DEVELOPED IN THE CORE OF A SMALL SYNCLINE

The cleavage planes are shearing planes in the upper part of the fold, and are parallel to the stretched limbs of minute puckers in the central part. In the lower sandy part, cleavage is absent. Mitta Mitta River, Victoria. (× 1 approx.)

plane of the strain ellipsoid, that it is necessarily parallel with the axial (apical) planes of folds, and conversely, that axial plane cleavage is flow cleavage. Exceptions can be quoted to all these rules, and the term is best used only to describe the texture and microstructure of a rock.

Fracture cleavage, on the other hand, was defined as 'conditioned by the existence of incipient, cemented, or welded parallel fractures and is independent of a parallel arrangement of the mineral constituents'.[1]

It is thus the antithesis of flow cleavage, although the latter

[1] Leith, C. K., *op. cit.*

is influenced by deformation-planes, and fracture cleavage grades to flow cleavage as the proportion of orientated mineral grains increases. The term fracture cleavage is applicable where the cleavage is determined by fractures along one or both of the conjugate shear directions in a deformed rock, but it has no constant relationship to the strain axes, since as shown by Harker and others, strong shearing on two directions may lead to cleavage in a composite plane between the two.[1] In rocks that exhibit pronounced shear fracturing, the presence of plastic matter such as clay plays an important part in binding the rock and probably too in facilitating overall flowage, as in argillaceous sandstones.

In slates and schists, a structure similar to fracture cleavage, but with marked flexing of the earlier cleavage or foliation along the shear planes, may be present. This is strain-slip cleavage, slip cleavage or shear cleavage,[2] and is typically found only in rocks that have undergone a certain degree of recrystallization—slates, phyllites, and schists. Where strongly developed, lamellar minerals come to lie in the shear zones in parallel alignment, and the further intensification of the structure has been shown to lead to transverse schistosity[3] and it is obvious from the flexing of micaceous minerals and foliation along the shears that the whole rock was involved in flowage, a good deal of which went on without actual rupture. In described examples such cleavage varies from parallelism with the axial planes of folds, to angles of 45° with them.

Although the relationships of cleavage to stress and strain conditions are variable, its attitude in folded beds shows certain regular relationships to the strata, which make its use in field work of great importance.

[1] Harker, A., *Metamorphism*: London, 1932, pp. 152–7. Sander, B., *Gefügekunde der Gesteine*: Vienna, 1930.

[2] Harker, A., 'On Slaty Cleavage and Allied Rock Structures': *Rept. Brit. Assoc. Adv. Sci.*, 1885 (1886), pp. 813–52; Dale, T. N., 'The Slate Belt of Eastern New York and Central Vermont': *19th Ann. Rept. U.S. Geol. Surv.*, Pt. 3, 1897–98, pp. 153–307; Mead, W. J., 'Folding, Rock Flowage, and Foliate Structures': *Journ. Geol.*, Vol. 36, 1940, pp. 1007–21.

[3] White, W. S., 'Cleavage in East-Central Vermont': *Trans. Amer. Geophys. Union*, Vol. 30, 1949, pp. 587–94.

Relationship of Cleavage to Folding.[1]—Cleavage that is genetically connected with folding may be recognized by a close systematic relationship between the cleavage and the geometry of individual folds. However, in many folded zones a *cleavage fan* is developed, the dip of the cleavage gradually

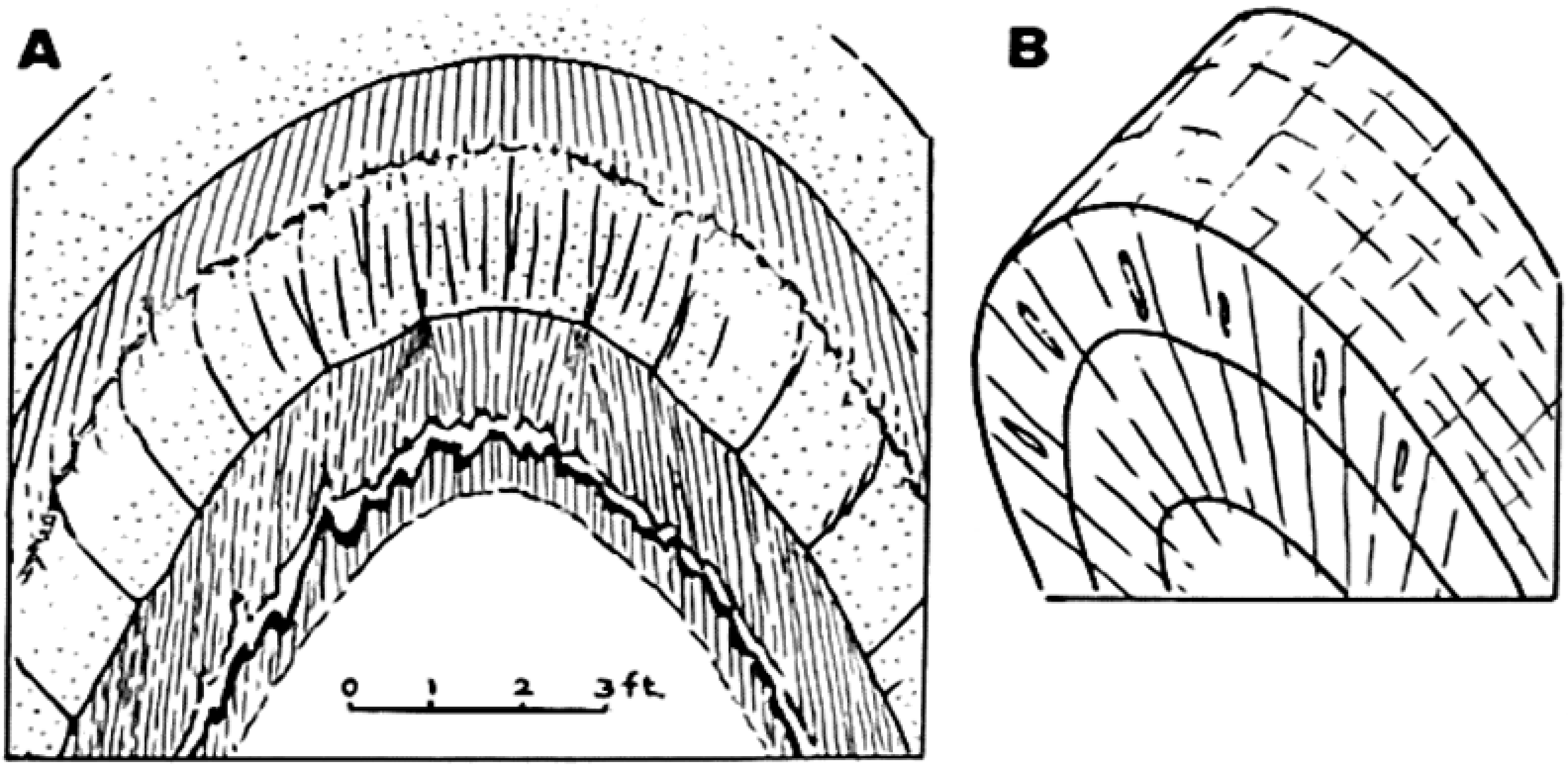


FIG. 67.—CLEAVAGE IN FOLDS

(After Fellows)

A. Cleavage dipping outwards in slate, and coarse cleavage ('fissuring') radiates inwards in sandstone, in graded beds. Note puckering of sandy laminae (black) in slate.

B. Radiating cleavage in anticline, with direction of elongation indicated by deformation of pebbles and ooids.

changing over considerable tracts of country across the strike, so that it converges either upwards or downwards. This arrangement, which is seen in Snowdonia, Denbighshire, and

[1] Cloos, H., and H. Martin, 'Der Gang einer Falte': *Fortschr. der Geol. u. Pal.*, Vol. 11, 1933, pp. 74–88. Scholtz, H., 'Faltung und Schieferung im Ostsäuerlander Hauptsattel': *Cbl. f. Min., &c.*, Abt. B, 1932, pp. 321–35. Dale, T. N., 'The Slate Belt of Eastern New York and Western Vermont': *19th Ann. Rept. U.S. Geol. Surv.*, Pt. 3, 1899, pp. 153–307. Scholtz, H., 'Über das Alter der Schieferung und ihr Verhaltnis zur Faltung': *Jahrb. d. Preuss. Geol. Landesanst.*, Vol. 52, 1931, pp. 303–16. Greenly, E., 'Foliation and its Relation to Folding in the Mona Complex at Rhoscolyn (Anglesey)': *Quart. Journ. Geol. Soc.*, Vol. 86, 1930, pp. 169–90. Gair, J. E., 'Cleavage and the Distortion of Stratigraphic Thicknesses in Appalachian Folds': *Trans. Amer. Geophys. Union*, Vol. 30, 1949, pp. 116–18. Engel, A. E. J., 'Studies of Cleavage, etc.': *Trans. Amer. Geophys. Union*, Vol. 30, 1949, pp. 767–84.

Flintshire[1] and also in the Rhineland[2] is clearly connected with the regional tectonics, but not directly with the folds themselves.

In dealing with slaty cleavage, which will be taken as typical, the relationship to folding varies. In folds having the characteristics of shear folds, the cleavage is strictly parallel with the apical plane of the fold (Fig. 55, A), but commonly it deviates from the apical plane in the limbs. It may dip uniformly either inwards or outwards or again outwards in the argillaceous beds and inwards in the arenaceous, in an anticline. Strong cleavage-zones in the sandstones have an appearance like fissures, and where these are closely spaced, as they are at fold axes, they may be mistaken for bedding planes. The puckering of

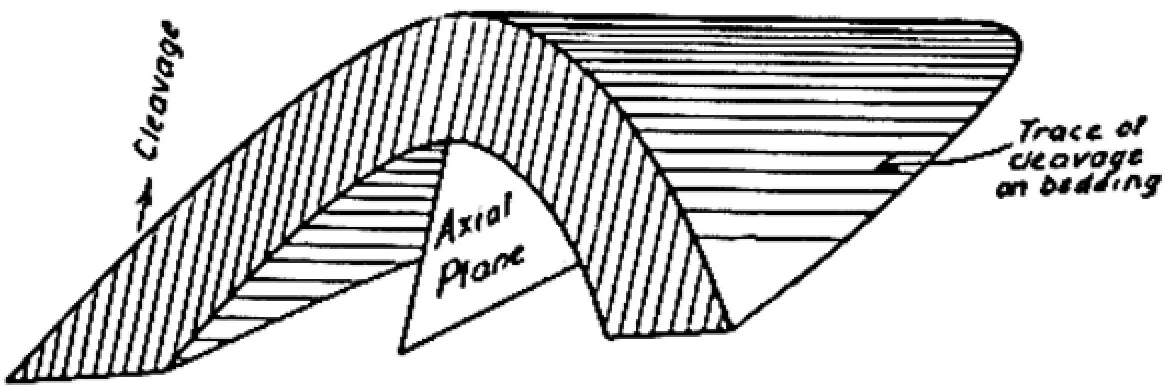


FIG. 68.—SHOWING CLEAVAGE PARALLEL TO THE AXIAL PLANE OF AN INCLINED PITCHING FOLD

The trace of the cleavage on the bedding indicates the pitch.

arenaceous or calcareous laminae in slates is characteristic, and is very strongly marked near the axes (Fig. 67).

E. Cloos' study of the South Mountain fold in Maryland[3] demonstrates by the study of deformed oolite grains in the limestones of the district that the variations in cleavage dip and intensity are directly related to the internal strains. Cleavage is parallel with the A–B planes of the oolite grains which may be regarded as strain ellipsoids. The results are of considerable significance in stratigraphy since they indicate that

[1] Boswell, P. G. H., *The Middle Silurian Rocks of North Wales*: London, 1949, pp. 96–108.

[2] Scholtz, H., 'Das varitische Bewegungsbild': *Fortschr. d. Geol. u. Pal.*, Hft. 25, 1930.

[3] 'Oolite Deformation in the South Mountain Fold, Maryland': *Bull. Geol. Soc. Amer.*, Vol. 58, 1947, pp. 843–918.

stratigraphic thicknesses measured normal to the bedding may be erroneous.

On the assumption that slaty cleavage is approximately parallel to the axial plane of folds, it may be used as a guide in the interpretation of complex fold structures. The pitch of a fold is indicated by the trace of the cleavage on the bedding planes in the fold limbs (see Fig. 68), because of the parallelism of the cleavage with the axial plane of the fold. Also, from an isolated outcrop showing cleavage and bedding, it can be

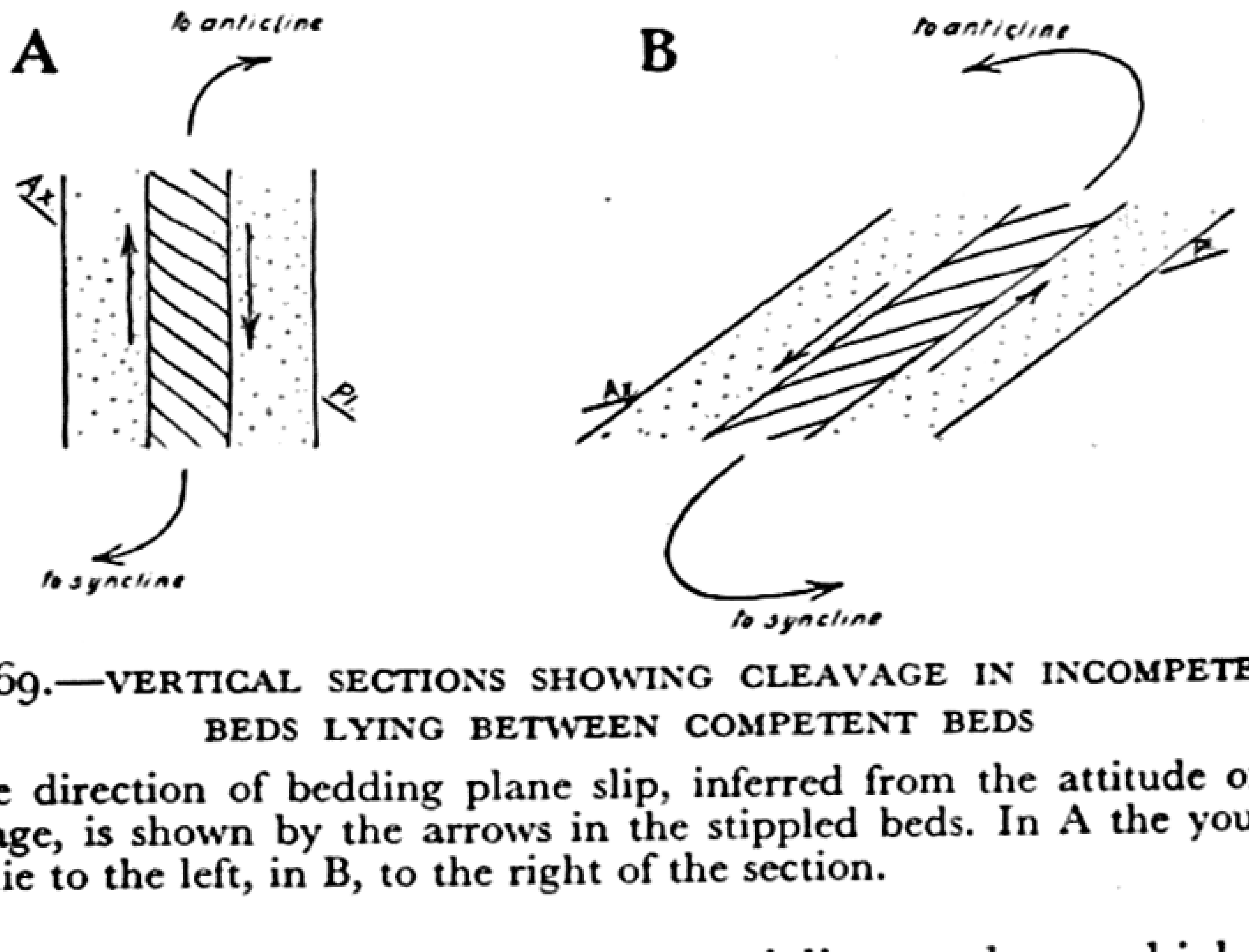


FIG. 69.—VERTICAL SECTIONS SHOWING CLEAVAGE IN INCOMPETENT BEDS LYING BETWEEN COMPETENT BEDS

The direction of bedding plane slip, inferred from the attitude of the cleavage, is shown by the arrows in the stippled beds. In A the younger beds lie to the left, in B, to the right of the section.

decided on which side the next anticline and on which the next syncline lies, if incompetent beds only are used in the test. The interpretation of competent beds is difficult and should not be relied upon. The rule is that the acute angle between bedding and cleavage points in the direction of bedding plane slip undergone by the adjacent bed (see Fig. 69), thus enabling the order of superposition to be determined, as with drag folding. Where the cleavage dips at a lower angle than the beds, overturning may at once be suspected.

Boudinage.—This is a structure in which beds set in a yielding matrix are divided by cross-fractures into pillow-like sectors. The cross-fractures are not sharp, but rather rounded and they

may be compared with the 'necks' that develop in ductile metal test pieces under tension. Boudinage is seen in limestones and sandstones interbedded with shales in fold limbs, and again in schists and gneisses.[1]

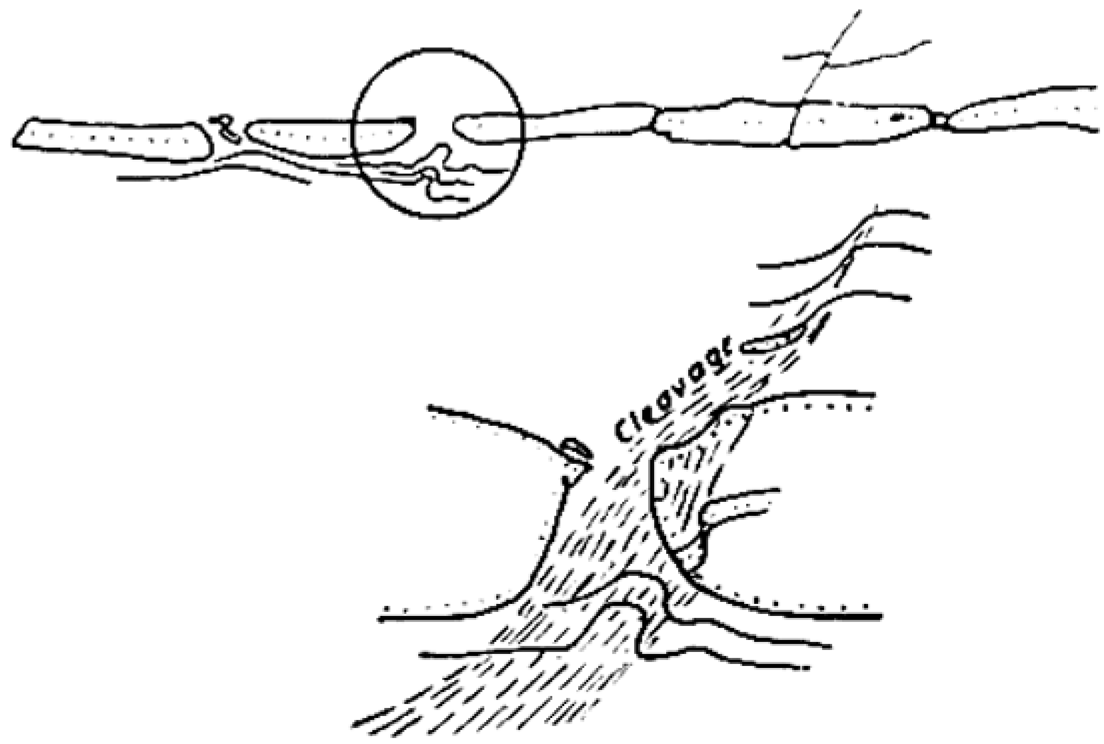


FIG. 70.—BOUDINAGE IN SANDSTONE

Sandstone interbedded in laminated slate, and showing boudinage due to stretching. Detail of the part encircled is shown below.

[1] Corin, F., 'A propos du boudinage en Ardenne': *Bull. Soc. Belge de Géol.*, Vol. 42, 1932, pp. 101–14. Wegmann, C. E., 'Note Sur le Boudinage': *Bull. Soc. Géol. France*, Ser. 5, Vol. 2, 1932, pp. 477–91. Walls, R., 'A new record of boudinage structure from Scotland': *Geol. Mag.*, Vol. 74, 1937, pp. 325–32.

Chapter V

FAULTS

A FAULT is a fracture along which observable displacement of crustal blocks has occurred, parallel to the plane of the break. Fault planes are therefore shearing planes, and their mechanical interpretation may be based upon the analyses of stress and strain relationships already given. In this chapter we shall be principally concerned with the description of the various kinds of faults and of the minor structures associated with them.

1. *NOMENCLATURE OF FAULTS*

Having regard to the direction of displacement of the crustal blocks relative to each other, we may classify faults as *dip-slip*, *strike-slip*, or *oblique-slip faults*. In dip-slip faults the direction of relative displacement of the blocks is parallel to the dip of the fault plane; in strike-slip faults it is parallel to the strike of the fault plane, and in oblique-slip faults, oblique along the fault plane.[1]

If the plane of a fault is not vertical, the *hanging wall* is that face of rock which lies above the fault plane, and the *foot wall* that which lies below. The *hade* is the angle between the fault plane and a vertical plane, and is thus the complement of the dip of the fault plane.

By an examination of the faulted strata or of the minor structures associated with the fault, such as slickensides and joints (see pp. 130–5), it is often possible to determine the

[1] 'Report of the Committee on the Nomenclature of Faults': *Bull. Geol. Soc. Amer.*, Vol. 24, 1913, pp. 163–86.

direction of relative displacement of the hanging and foot walls. In using the term 'relative displacement' nothing is implied as to which block actually moved; a downward relative displacement of the hanging wall block may occur by actual downthrow of this block, by upthrow of the foot wall block, or by both these mechanisms. As is shown in Fig. 71, it is not possible to determine even the relative displacement of the

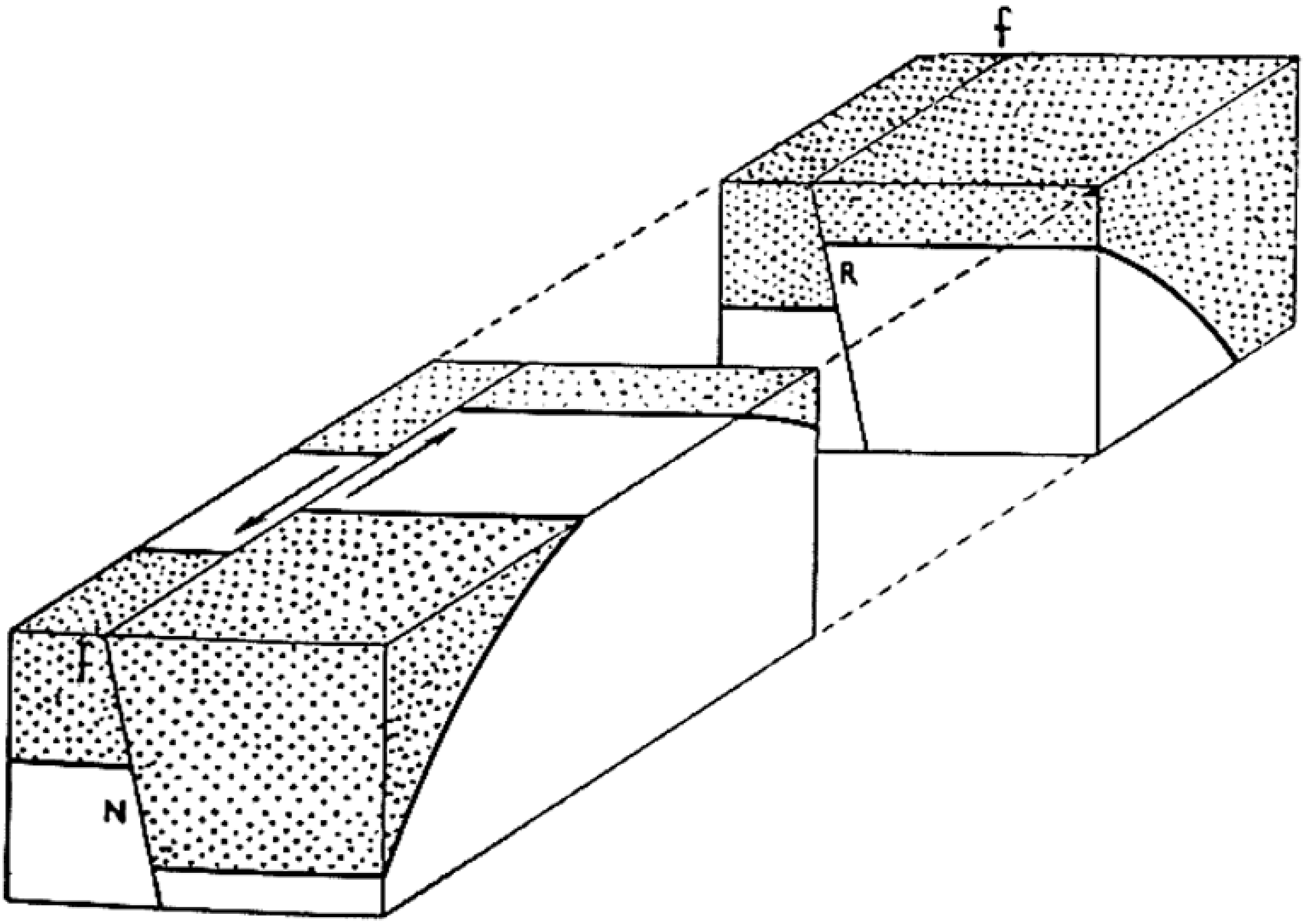


FIG. 71.—BLOCK DIAGRAM OF A STRIKE-SLIP DIP FAULT (*f*–*f*) AFFECTING AN ANTICLINE

(After Gill, 1935)

At N the fault appears in section to be normal, and at R reverse.

blocks from inspection of the faulted beds along one cross-section only, because the position of the beds in the hanging wall relative to those in the foot wall along this line will depend not only on the displacement due to the fault, but also on the attitude of the fault plane and of the beds.

Faults in which the displacement of the hanging wall is downwards relatively to the foot wall are called *normal faults*, and *reverse faults* are those in which the displacement of the

hanging wall relative to the foot wall is upwards.[1] A dip-slip fault in which the hanging wall is displaced downwards relatively to the foot wall is therefore described as a *dip-slip normal fault*. We may also have dip-slip reverse faults and either normal or reverse oblique-slip faults, but the terms normal and reverse obviously cannot be applied to strike-slip faults, for in these there is no relative movements in a vertical direction.

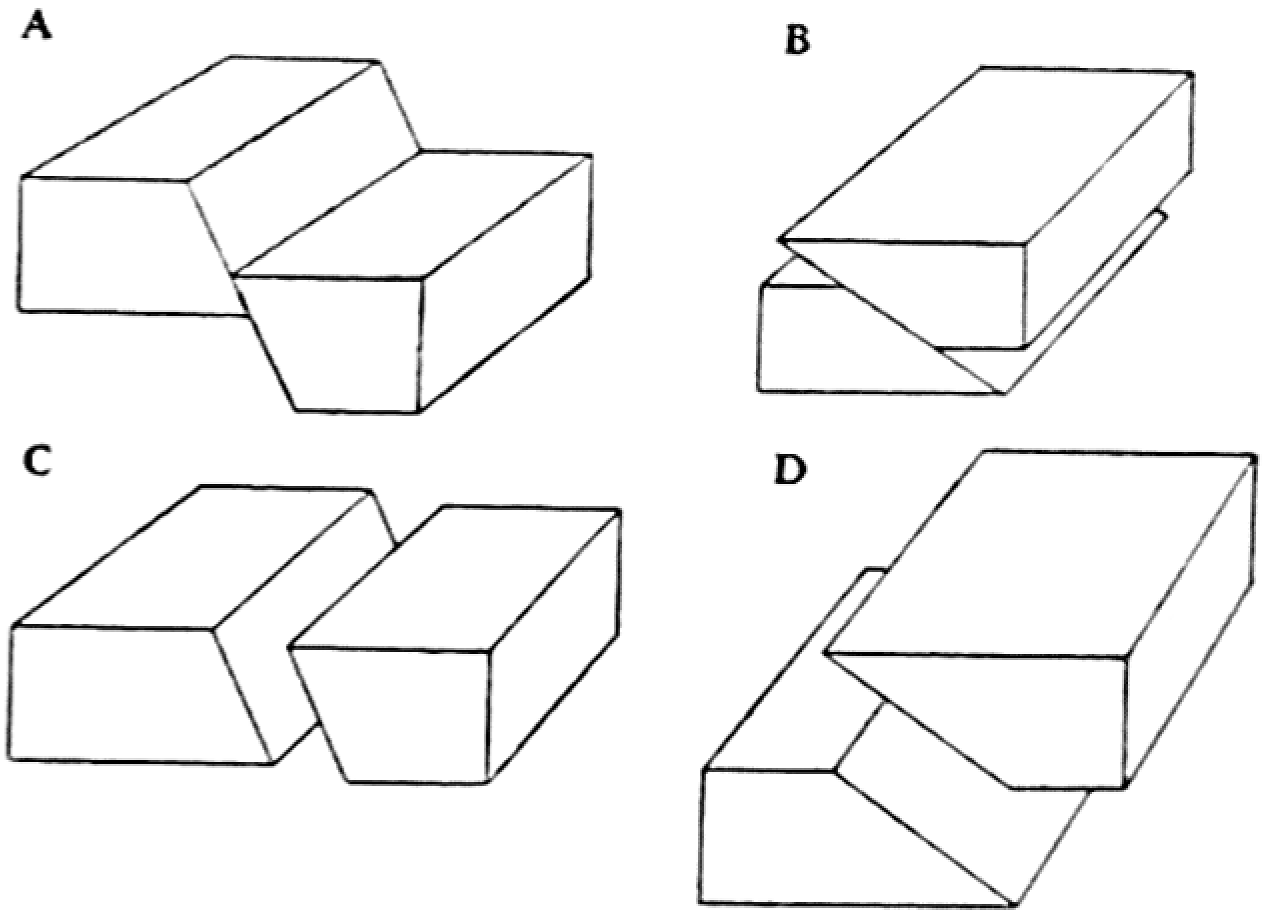


FIG. 72

A, dip-slip normal fault; B, dip-slip reverse fault; C, oblique-slip normal fault; D, oblique-slip reverse fault.

It should be noted that the Committee on Fault Nomenclature, instituted by the Geological Society of America, recommended in 1913[2] that a fault should be termed normal if the *apparent* displacement of a particular bed in the hanging wall along a given line of section is downwards relatively to the same bed in the foot wall. An apparently normal displacement does not, however, necessarily mean that the actual relative

[1] Equivalents for these terms in German and French are: for normal faults, *Abschiebungen*, *Verwerfungen*, *failles normales*, and for reverse faults, *Aufschiebungen*, *Überschiebungen*, *failles inverses*.

[2] *Bull. Geol. Soc. Amer.*, Vol. 24, 1913, pp. 163–86.

movement of the hanging wall block was downwards (see Fig. 71), and since the Committee's usage has been adhered to by some American authors, it is obvious that, if confusion is to be avoided, care must be exercised in reading.

If we consider the relationships between the trace of the fault plane at the surface and the attitude of the faulted beds, a fault may be described as a *strike fault* if the fault plane strikes in the same general direction as the beds, a *dip fault* if it strikes in the general direction of the dip of the beds, and an *oblique fault* if it is markedly oblique to the strike of the beds. If the fault plane coincides with a bedding plane, the fault is called a *bedding fault*. Such faults are not easily recognizable in the field, although they are quite common, and curved fault planes may frequently be traced into bedding planes, where the relative displacement is not visible.

The nomenclature of the chief types of faults may be summarized as follows (see also Fig. 72):

Direction of movement in the fault plane—	Displacement of hanging wall relative to foot wall—	Strike of fault plane relative to strike and dip of beds—
dip-slip	normal	strike fault
oblique-slip	reverse	dip fault
strike-slip		oblique fault

In this table any required term from one column may be combined with any required term from the others to give the full descriptive terminology of a particular fault, e.g. a normal fault in which the movement is oblique along the fault plane and in which the strike of the fault is at right angles to that of the faulted beds is an *oblique-slip normal dip fault.*

2. *NOMENCLATURE OF FAULT DISPLACEMENTS*[1]

Slip and Shift.—*Slip* refers to the relative displacement of formerly adjacent points, measured along the fault plane. The *net-slip* is the resultant of the *strike-slip* and the *dip-slip*, which refer to the displacement parallel to the strike and to the dip of the fault plane respectively (see Fig. 73). The displacement of crustal blocks is frequently not restricted to movements along one fault plane, but takes place in part by flexing

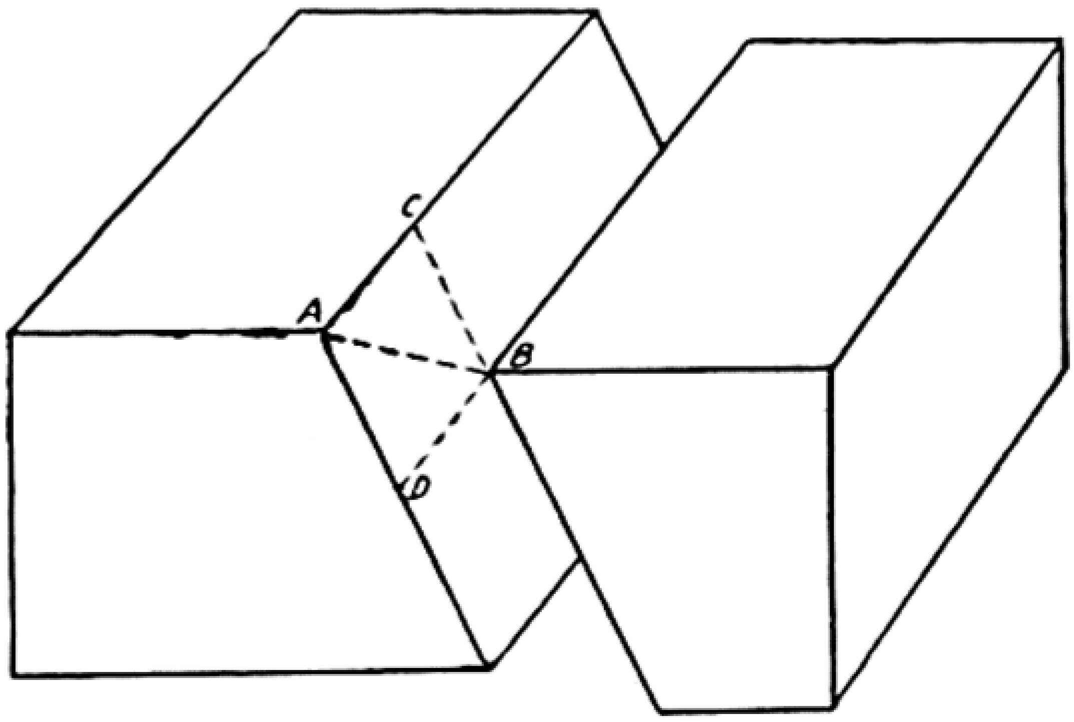


FIG. 73.—AN OBLIQUE-SLIP NORMAL FAULT
AB, net-slip; AC, strike-slip; AD, dip-slip.

adjacent to the fault (see Fig. 74 and p. 119), or by means of slip along closely spaced parallel shearing planes. The zone of disturbed rocks between the faulted blocks is termed the *fault zone*, and, as may be seen from Fig. 74, in describing the movement of the blocks as a whole, it is necessary to consider the relative displacement of points lying outside this zone. The term *shift* is used to denote the relative displacement of points far enough removed from the fault to be unaffected by local disturbance in the fault zone, and in broad discussions of

[1] For further information, consult the 'Report of the Committee on the Nomenclature of Faults', *Bull. Geol. Soc. Amer.*, Vol. 24, 1913, pp. 163–86. Gill, J. E., 'Normal and Reverse Faults': *Journ. Geol.*, Vol. 43, 1935, pp. 1071–9. Straley, H. W., III, 'Some Notes on the Nomenclature of Faults': *Journ. Geol.*, Vol. 42, 1934, pp. 756–63. Challinor, J., 'The "Throw" of a Fault': *Geol. Mag.*, Vol. 70, 1933, pp. 385–93; 'The Primary and Secondary Elements of a Fault': *Proc. Geol. Assoc.*, Vol. 57, 1946, pp. 153–60.

faulting, it is more important than the *slip*. More specifically we can refer to the *net-shift*, *dip-shift*, and *strike* or *lateral shift*.

Separation.—This refers to the apparent displacement of the two comparable parts of a faulted bed or lode (e.g. top to top or bottom to bottom), measured in cross-sections taken in any required direction. The *normal horizontal separation*, or *offset*, is the distance between the displaced portions of the faulted body measured at right angles to its strike, in a horizontal plane

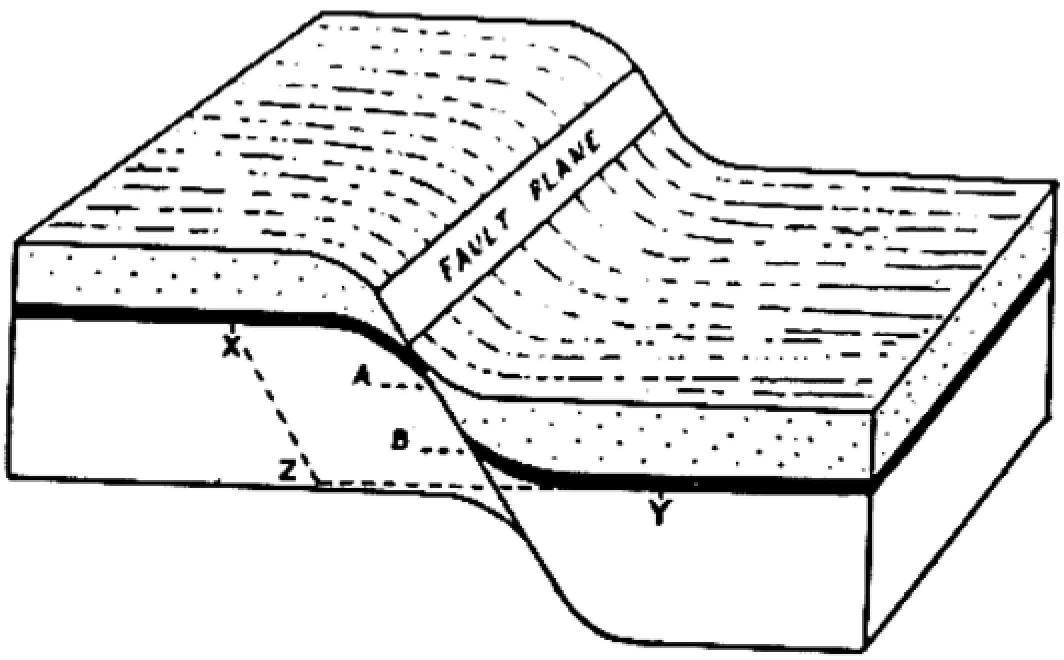


FIG. 74.—A DIP-SLIP NORMAL FAULT, SHOWING FLEXING OF THE BEDS NEAR THE FAULT PLANE

AB is the dip-slip; XZ is the dip-shift.

(Fig. 75). The *vertical separation* is often important in mining, and is the vertical distance between comparable parts of the faulted body. The *dip separation*[1] is the distance between comparable parts of the faulted body on the hanging and foot walls respectively, measured parallel to the dip of the fault plane. If the intersection of the body with the fault plane on the hanging wall lies below that on the foot wall in a given cross-section, the faulted body shows a *normal dip separation* in the plane of the section. If the reverse is the case, it shows a *reverse dip separation*. *Throw* and *heave* are now used exclusively for apparent displacements as seen in a cross-section normal to the fault plane, not for the movement itself. *Throw* is the vertical distance separating the faulted parts of a bed, and *heave* the horizontal distance.

The possible geometrical relationships of strata lying in

[1] Gill, J. E., *op. cit.*, 1935.

different attitudes and faulted in various ways are too numerous to permit adequate discussion in a book of this nature. For more detailed treatment of this subject the works mentioned below may be consulted.[1]

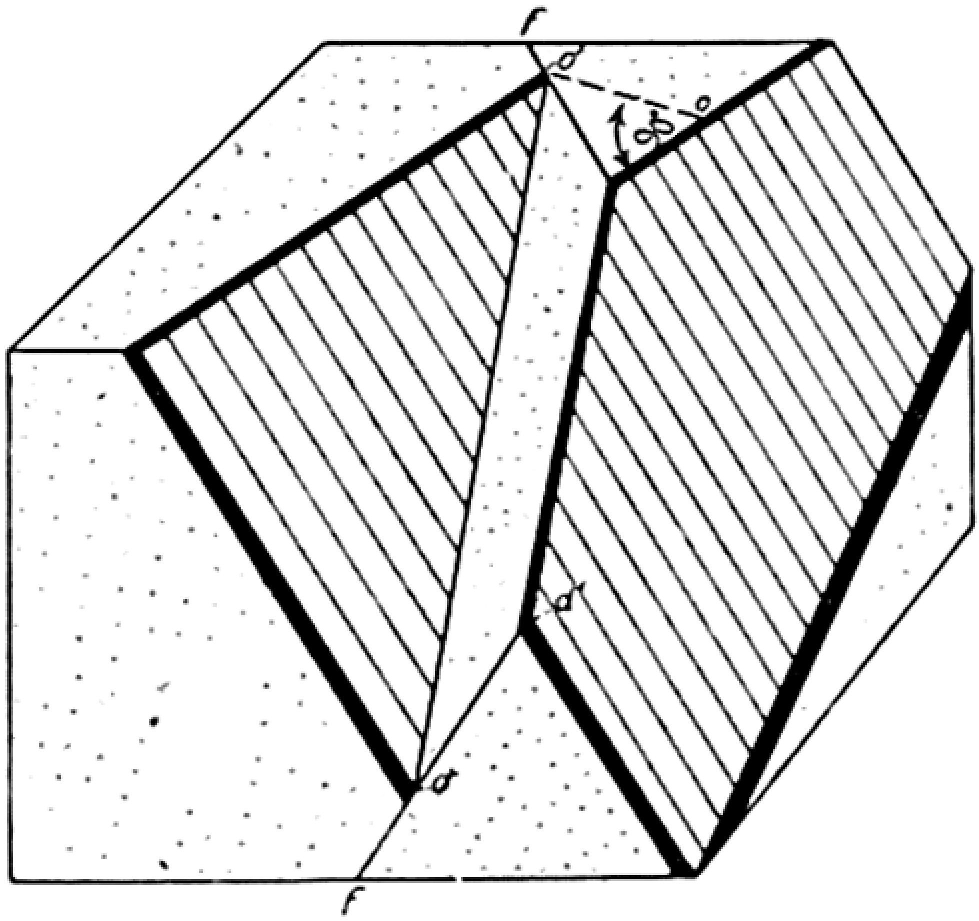


FIG. 75.—BLOCK-DIAGRAM OF A FAULTED BED (BLACK), SHOWING THE OFFSET (OO′) AND THE APPARENT DIP SEPARATION (DD′)

3. *DISCUSSION OF FAULTING*

Normal Faults.—In normal faults that are caused by regional crustal tension (Fig. 41), or which are associated with the sagging under gravity of portions of the crust from which support, e.g. in the form of underlying magma, has been removed (*gravity faults*), the hanging wall moves downwards relatively to the foot wall. Actual upthrust of the foot wall takes place, however, in those normal faults that are caused by vertical push from beneath (Fig. 54), and in some *cylindrical faults*. A cylindrical fault involves the rotation of a crustal block about an axis, parallel to a fault plane that has the curvature of portion of a cylinder. A normal fault formed at the surface by upward rotation of the foot wall along such a cylindrical

[1] Earle, K. W., *Dip and Strike Problems*: London, 1934. Brown, B. C., and F. Debenham, *Sructure and Surface*: London, 1929. Tolman, C. F., *Graphical Solution of Fault Problems*: London, 1911. Haddock. M. H., *Disrupted Strata*, London, 2nd edn., 1938.

fault plane passes into a high angle reverse fault in depth[1] (see Fig. 76), and it is therefore clear that the distinction between normal and reverse faulting is not always of fundamental significance dynamically.

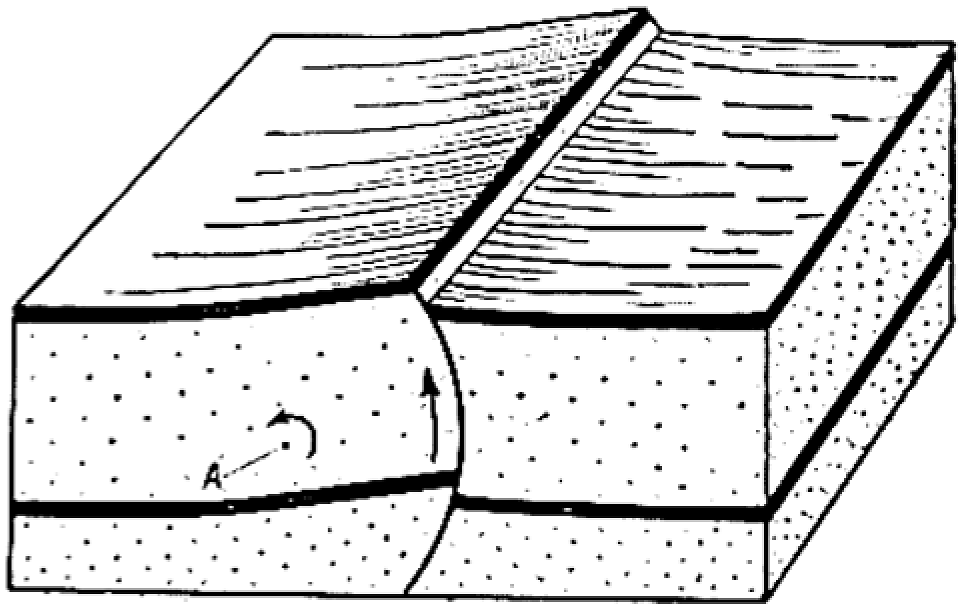


FIG. 76.—BLOCK DIAGRAM OF A CYLINDRICAL FAULT
A, axis of rotation.

An individual normal fault may pass either laterally or vertically into a monoclinal flexure, or it may retain its fault characteristics throughout its entire length, and die out by diminution of the throw. In the latter case the fault is generally crescentic in plan.[2]

Pivotal Faults.—In pivotal faulting the block that is depressed at one locality is elevated at another (Fig. 77). This may be due

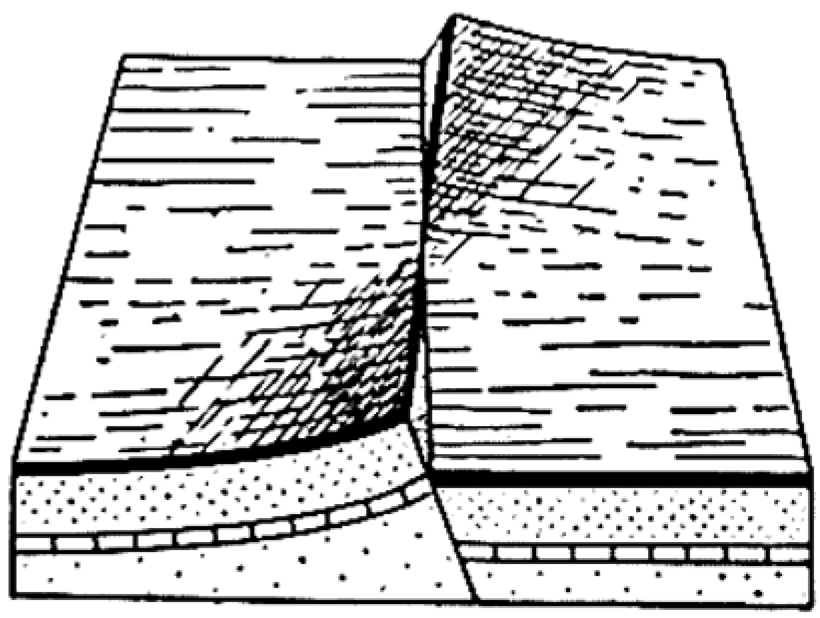


FIG. 77.—BLOCK DIAGRAM SHOWING REVERSAL OF UP-AND DOWN-THROWN SIDES ALONG A PIVOTAL FAULT
(Adapted from Busk, *Earth Flexures*)

[1] Tolman, C. F., Jr., *Graphical Solution of Fault Problems*: London, 1911, pp. 36, 37. Cloos, H., *Einführung in die Geologie*: Berlin, 1936, pp. 198–200.
[2] Busk, H. G., *Earth Flexures*: Cambridge, 1929, pp. 96–8.

to warping of the blocks, in which case the direction of dip of the fault plane changes where the reversal of the upthrown and downthrown sides takes place, or it may be due to rotation of a block about an axis perpendicular to the fault plane, in which case the fault is normal at one end and reverse at the other, the fault plane dipping in the same direction throughout (Fig. 78, A). In a *hinge fault* one of the blocks hinges about an axis at right angles to the fault line, the throw increasing away from the hinge (Fig. 78, B).

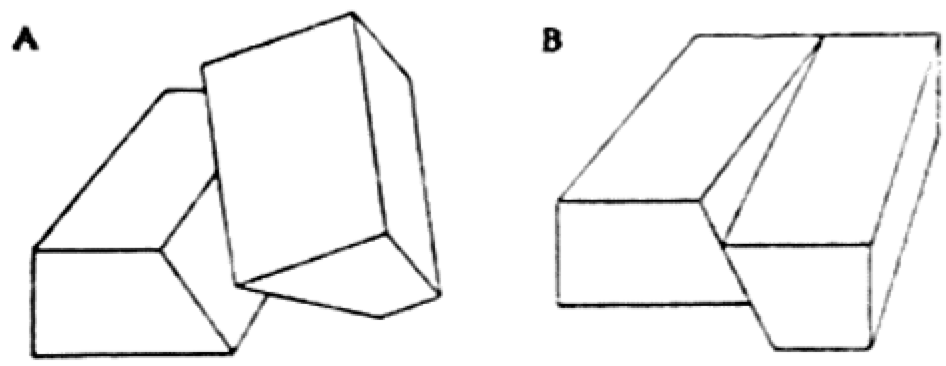


FIG. 78
A, a Pivotal fault; B, a Hinge fault.

Strike-slip Faults.—Many of the faults in active regions of the crust are of the strike-slip type. The San Andreas 'Rift', in California, which extends in an almost straight line for over 600 miles and has been active since early Tertiary times, is a well-known example. Displacements during recent earthquakes, as well as geological investigations, show clearly that the movements that have taken place along this fault are essentially horizontal. They are probably caused by shearing stress developed as a component of north–south compression acting in a horizontal plane.[1]

The Great Glen fault in Scotland is also a strike-slip fault, and it is estimated from the matching of rock masses on either side that the lateral slip may be approximately 65 miles.[2]

[1] Cloos, H., 'Bau und Bewegung der Gebirge in Nordamerika, Skandinavien und Mitteleuropa': *Fortsch. d. Geol. u. Pal.*, Vol. 7, Hft. 21, 1928, pp. 253–4. Bucher, W. H., *The Deformation of the Earth's Crust*: Princeton, 1933, pp. 314–21. Vickery, F. P., 'The Structural Dynamics of the Livermore Region': *Journ. Geol.*, Vol. 33, 1925, pp. 608–28. Willis, B., 'San Andreas Rift, California': *Journ. Geol.*, Vol. 46, 1938, pp. 793–827.

[2] Kennedy, W. Q., 'The Great Glen Fault': *Quart. Journ. Geol. Soc.*, Vol. 102, 1946, pp. 41–76.

Unequal advance of adjacent sectors of moving nappes or of normally folded strata results in the development, concurrently with the major structures, of strike-slip faults that trend transverse to the strike of the deformed rocks. Such faults are best termed *tear faults* (*transcurrent* or *transverse thrusts* of Geikie; *Blattverschiebungen*).[1] In general, only one of the two possible

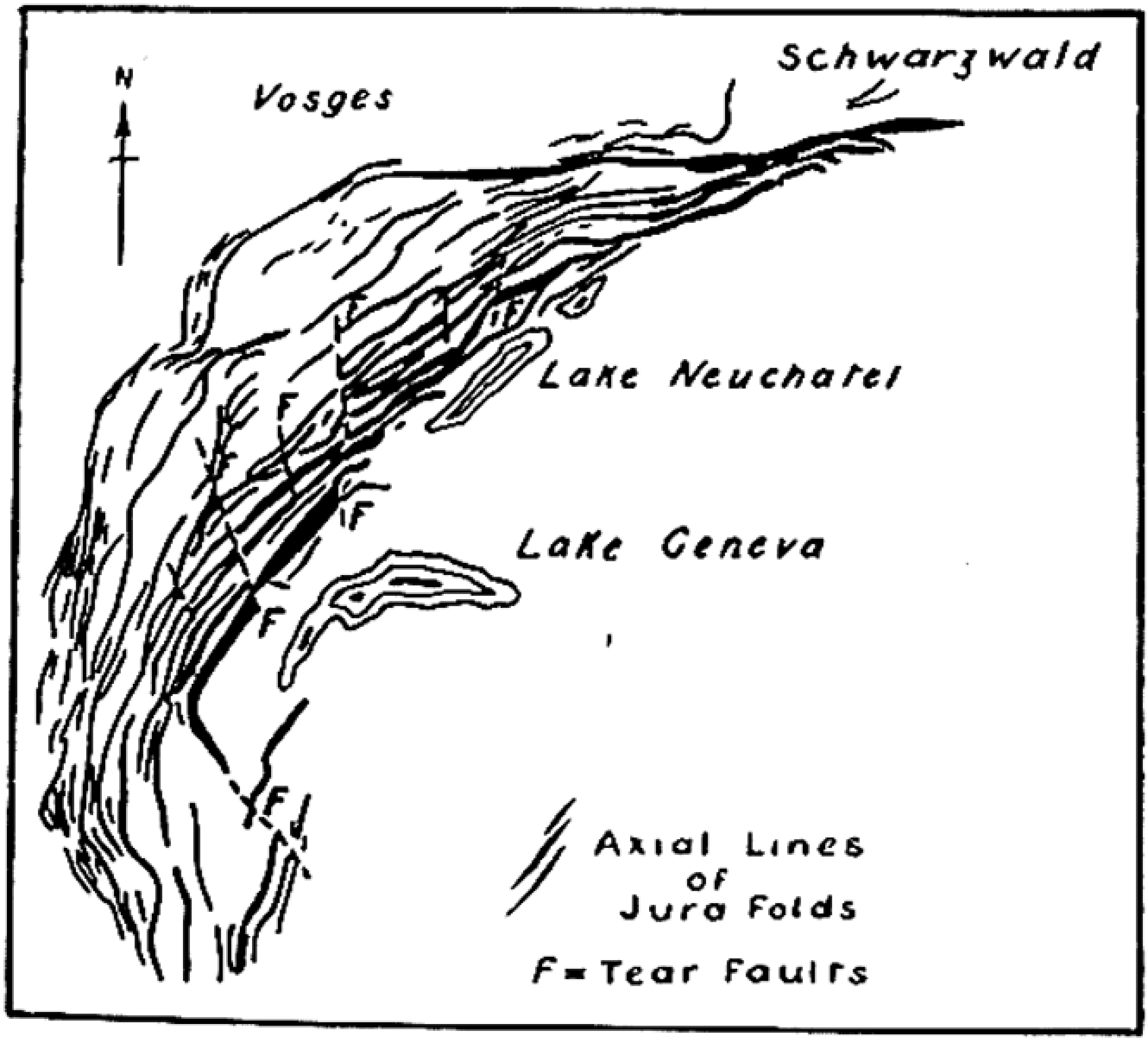


FIG. 79.—SKETCH MAP OF THE FOLDED JURAS, SHOWING THE CHIEF FOLDS AND THE TEAR FAULTS (F) ALONG WHICH THE FOLD AXES ARE OFFSET

(After Heim, from Bucher, *The Deformation of the Earth's Crust*)

directions along which shearing may take place between the moving blocks is strongly developed, the dominant direction being that which is most nearly parallel to the direction of movement of the rocks (see also p. 39). Important tear faults

[1] Geikie, J., *Structural and Field Geology*: New York, 3rd edn., 1912, pp. 179–80. Marr, J. E., *Quart. Journ. Geol. Soc.*, Vol. 62, 1906, pp. lxxvi–viii. See also Perry, E. L., 'Flaws and Tear Faults': *Amer. Journ. Sci.*, Vol. 29, 1935, pp. 112–24, for a critical discussion of the nomenclature of strike-slip faults, with full bibliography.

have been recognized in many regions of nappe structure and also of normal folding[1] (Fig. 79). They have been experimentally reproduced by Cloos and by Lee[2] (see pp. 62–3; Fig. 37, and Plate IV).

Anderson has classified strike-slip faults as either sinistral or dextral, according to the relative movements of the blocks. If the fault is observed from one side, it is sinistral if the displacement on the distant side is towards the left and dextral if towards the right.

The term transcurrent fault has been applied as a synonym for strike-slip fault, but its use is inappropriate for faults such as the San Andreas, that are rather parallel with than transverse to the regional structures. Such major faults, too, have tectonic relationships of an order different from the tear faults associated with folding, and the term *wrench fault* may be suggested as appropriate for them.

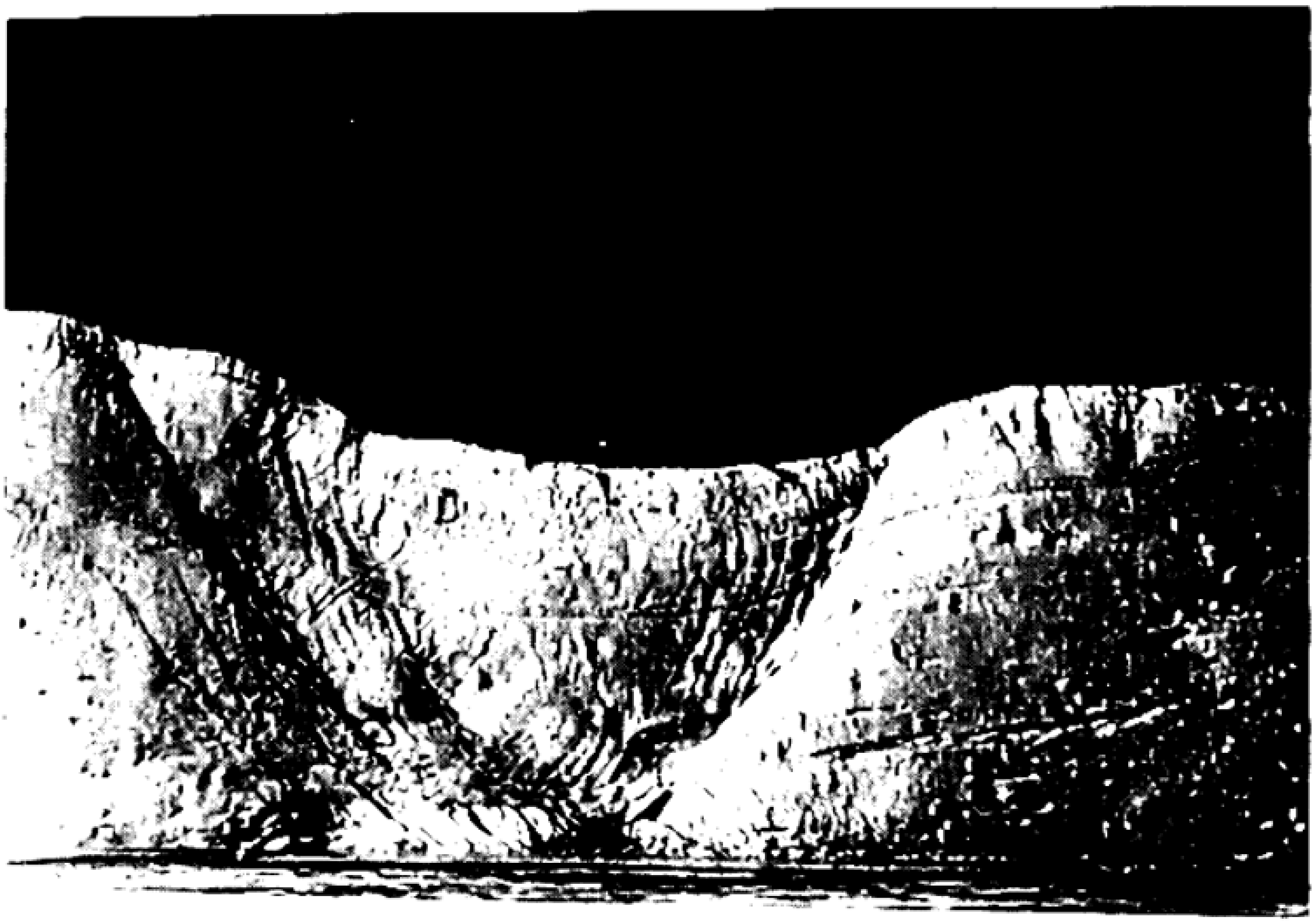
Reverse Faults.—A reverse fault involves a shortening of the section of faulted rocks, whereas a normal fault involves a lengthening. If the plane of a reverse fault dips at an angle of 45° or more, the fault is sometimes called an *upthrust*: if it dips at less than 45°, it is called an *overthrust* if the hanging wall is believed to have been actually thrust over the foot wall, or an *underthrust* if the foot wall is believed to have been pushed under the hanging wall.[3] Reverse faults in which the fault plane is horizontal, or nearly so, are called *low angle thrusts*, or if the translation of the thrust blocks is very great, *thrust nappes*.

Stretch thrusts are reverse faults which develop by the shearing ing through of the attenuated middle limbs of overturned folds. We have noted the existence of minute stretch thrusts in connexion with cleavage, but they may also occur on a grand scale as nappes (Fig. 32). Where a resistant stratum shears at a flexure, one limb is pushed over the other along a *break thrust*.

[1] Heim, A., *Geologie der Schweiz*, Vol. 1, 1919, p. 615, Fig. 103 (Juras).

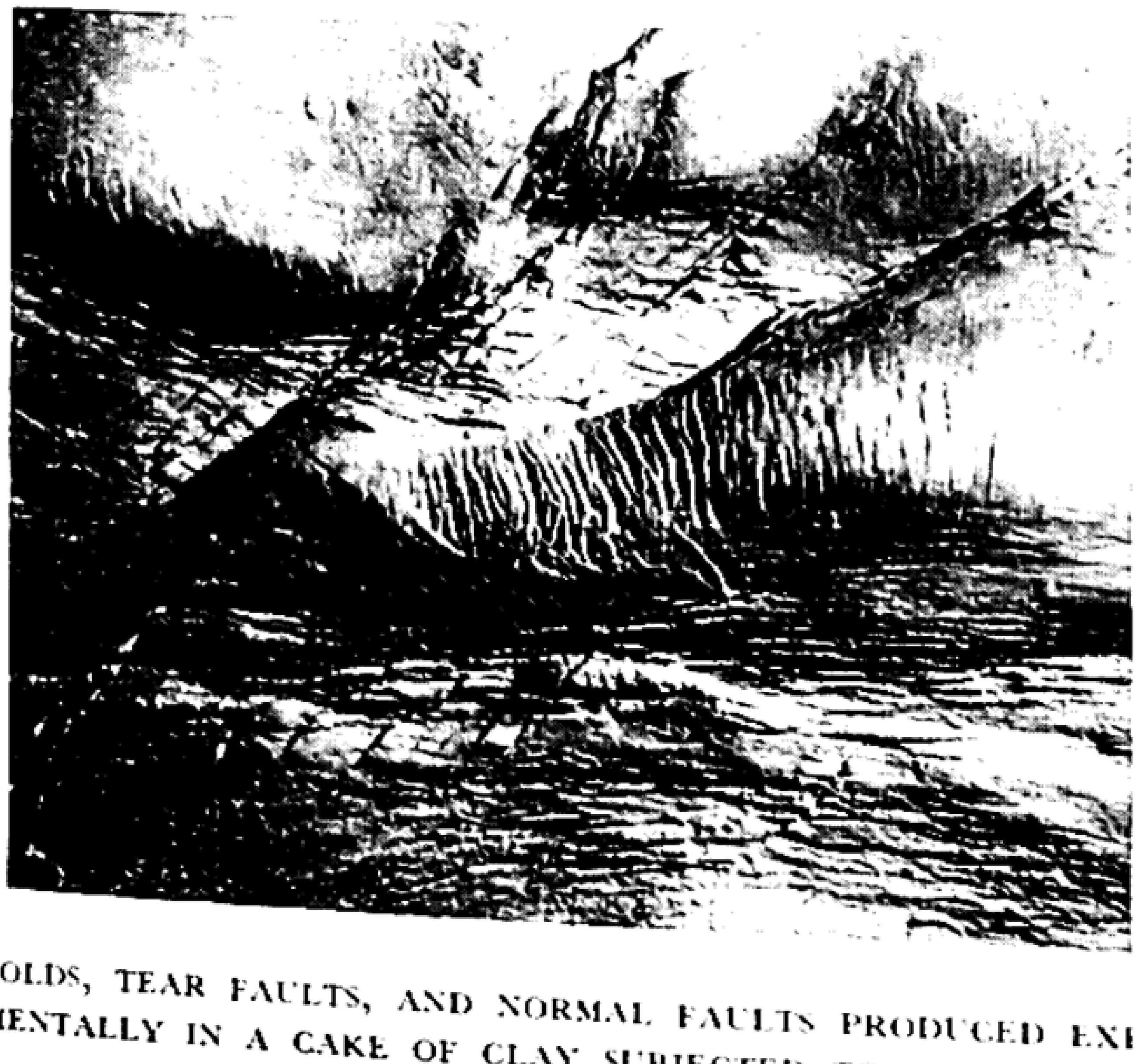
[2] Cloos, H., 'Zur Tektonischen Stellung des Saargebietes': *Zeitsch. d. Deutsch. Geol. Ges.*, Vol. 85, 1933, pp. 307–15. Lee, J. S., 'Some Characteristic Structural Types in Eastern Asia, &c.': *Geol. Mag.*, Vol. 66, 1929, *passim*.

[3] Lovering, T. S., 'Field Evidence to distinguish Overthrusting from Underthrusting': *Journ. Geol.*, Vol. 40, 1932, pp. 651–63.



A. MINIATURE GRABEN PRODUCED EXPERIMENTALLY IN A CAKE OF CLAY SUBJECTED TO TENSION

(After H. Cloos)



B. FOLDS, TEAR FAULTS, AND NORMAL FAULTS PRODUCED EXPERIMENTALLY IN A CAKE OF CLAY SUBJECTED TO COMPRESSION

(After H. Cloos)

These are generally low angle thrusts of restricted downward extension, the shortening of the section of rocks beneath the thrust being taken up as crumpling [1] (Fig. 80). *Shear thrusts* are

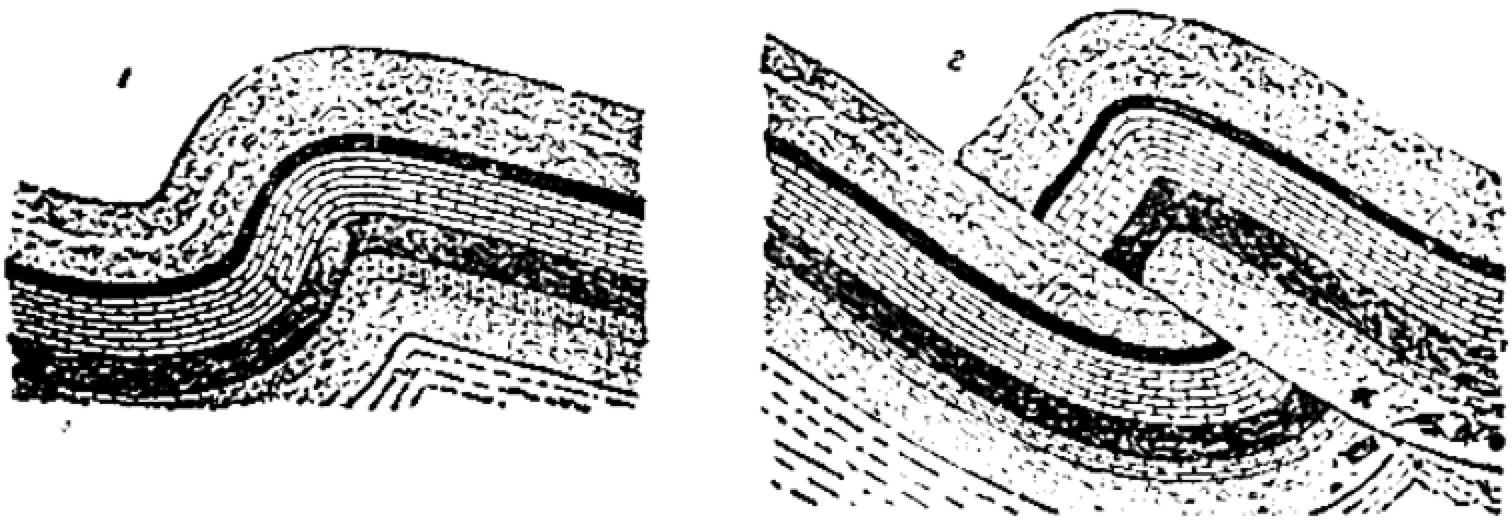


FIG. 80.—TWO STAGES IN THE DEVELOPMENT OF A BREAK THRUST
(After Willis, 1893)

shearing planes whose attitude is controlled by the mechanics of deformation and the physical properties of the rocks. They may develop in folded rocks by the squeezing of the beds in the

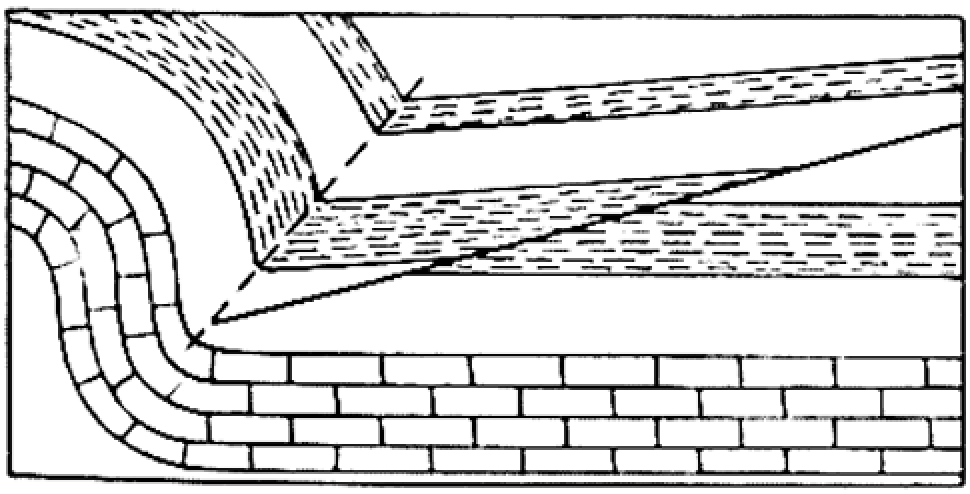


FIG. 81.—A SHEAR THRUST DEVELOPED IN WEAK ROCKS BY SQUEEZING IN FRONT OF A MONOCLINE IN LIMESTONE
(After R. Willis, 1935)

fold cores or in front of monoclinal flexures (see Fig. 81),[2] or they may have no connexion with folding (Fig. 82).

Most thrust faults along which very large displacement of

[1] Willis, R., 'Development of Thrust Faults': *Bull. Geol. Soc. Amer.*, Vol. 46, 1935, pp. 409–24. Billings, M., 'Thrusting Younger Rocks over Older': *Amer. Journ. Sci.*, Vol. 25, 1933, pp. 140–65. Willis, B., 'Mechanics of Appalachian Structure': *13th Ann. Rept. U.S. Geol. Surv.*, Pt. 2, 1893, pp. 222–3.

[2] Willis, R., 'Development of Thrust Faults': *Bull. Geol. Soc. Amer.*, Vol. 46, 1935, pp. 409–24.

the thrust blocks has occurred are low angle thrusts. Some of these are true shear thrusts, and are probably caused by shearing stress acting in a vertical plane.[1] The experiments of Cadell,[2] designed to imitate the structure of the north-west Highlands of Scotland, showed, however, that the application of direct compression, after first causing the piling up of a number of slices of rock by thrusts dipping at about 45°, finally produces low angle thrusts along which the piled-up slices move bodily (see Fig. 83). This may perhaps be due to the development of rotational strain within the deformed material,[3] due, for instance, to a decrease in the resistance of the superficial portions relatively to those lying beneath, which would

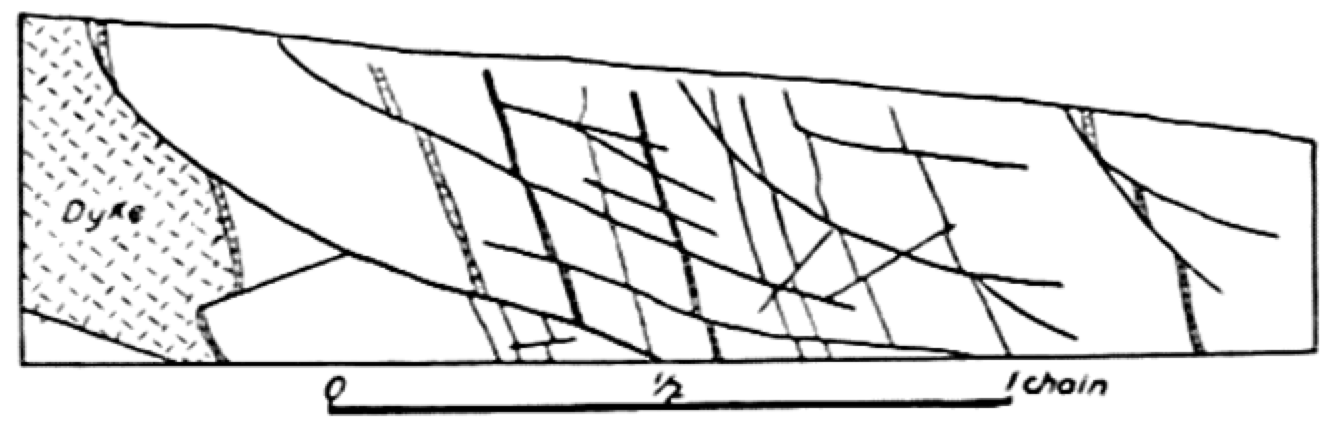


FIG. 82.—SHEAR THRUSTS, EACH OF WHICH SHOWS VARIATION IN THE AMOUNT OF SLIP FROM POINT TO POINT, CUTTING HIGH-DIPPING SANDSTONES AND MUDSTONES AT STUDLEY PARK, VICTORIA

permit more rapid advance of the upper parts. Kaisin[4] has emphasized for the Ardennes the importance of differences in the rate of advance of masses separated by faults. Such differences arise in rocks undergoing general flowage, because of differences in physical properties from place to place, and because of the freedom from constraint of the superficial masses. A distinction may be made between first-order thrust

[1] Chamberlin, R. T., and W. Z. Miller, 'Low Angle Faulting': *Journ. Geol.*, Vol. 26, 1918, pp. 1–44. Quirke, T. T., 'Concerning the Process of Thrust Faulting': *Journ. Geol.*, Vol. 28, 1920, pp. 417–38.

[2] Cadell, H. M., 'Experimental Researches in Mountain Building': *Trans. Roy. Soc. Edinburgh*, Vol. 35, 1890, pp. 337–57.

[3] Chamberlin, R. T., and W. Z. Miller, 'Low Angle Faulting': *Journ. Geol.*, Vol. 26, 1918, pp. 1–44.

[4] Kaisin, F., 'Poussées tangentielles ou "Champs tectoniques"?': *Bull. Soc. Belge de Géol.*, Vol. 53, 1944, pp. 228–63.

faults (*failles de charriage*) which die away upwards, and away from the zone in which the structures have their origin (as with the basal thrust planes in Fig. 83), and second-order faults (*failles d'entrainement*), which originate at major thrusts and die out away from these (i.e downwards, beneath the main fault).

Imbricate structure (*Schuppenstruktur*), well exhibited in the north-west Highlands of Scotland, consists of a series of thin rock slices separated by high angle thrusts. In the Glencoul area, the zone of imbricate structure is bounded above and below by low angle thrusts of later development, and the thrusts in the imbricate zone are unconnected with folds, being true shear thrusts[1] (see Fig. 29, p. 54).

FIG. 83.—EXPERIMENTAL REPRODUCTION OF IMBRICATE STRUCTURE
(After Cadell, 1890)

Pressure was applied in the direction of the arrow, producing first the imbricate structure and then the basal thrust plane or *sole*, T.P.

If a plane along which thrusting is taking place reaches the earth's surface, over which the thrust block then advances, the fault is termed a *surface thrust* (*Reliefüberschiebung*).[2] Such faults may arise by erosion of the arch of a major anticline, which allows a competent member in one limb to slip over underlying incompetent beds, and to move forwards over the erosion surface. They are then called *erosion thrusts*,[3] being dependent upon the erosion of the anticline (Fig. 84). Busk has described erosion thrusts in Persia which are not due to the

[1] Peach, B. N., and J. Horne, *Guide to the Geological Model of the Assynt Mountains*: Edinburgh, 1914, pp. 17–19.

[2] Ampferer, O., 'Beiträge zur Auflösung der Mechanik der Alpen': *Jahrb. Geol. Bundesanst., Wien*, Vols. 73–81, 1923–30.

[3] Willis, B., 'The Mechanics of Appalachian Structure': *13th Ann. Rept. U.S. Geol. Surv.*, Pt. 2, 1893, pp. 222–3.

sliding of the competent beds over the incompetent, but to the squeezing out of mobile gypsum beds through the eroded crests and limbs of anticlines, or at other places where the cover of

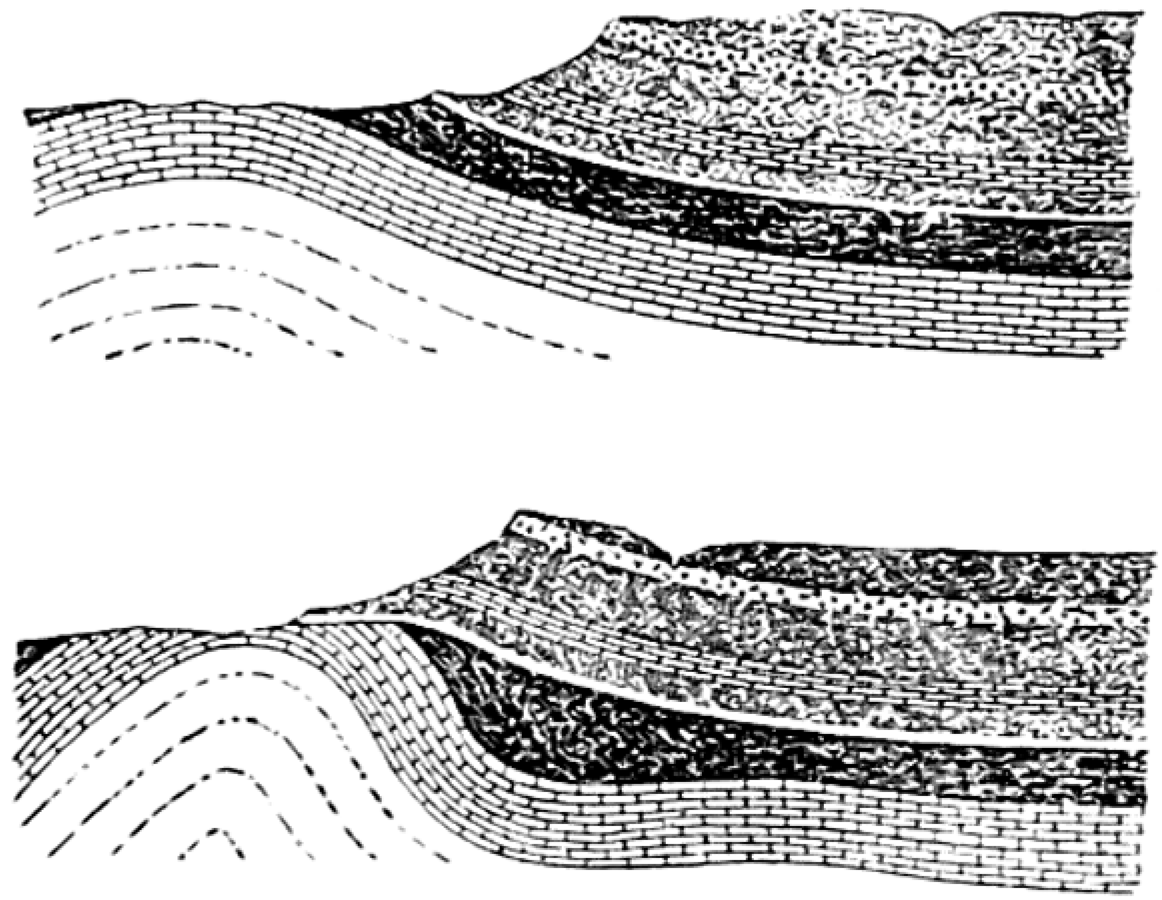


FIG. 84.—TWO STAGES IN THE DEVELOPMENT OF AN EROSION THRUST AT AN ERODED ANTICLINE

(After Willis, 1893)

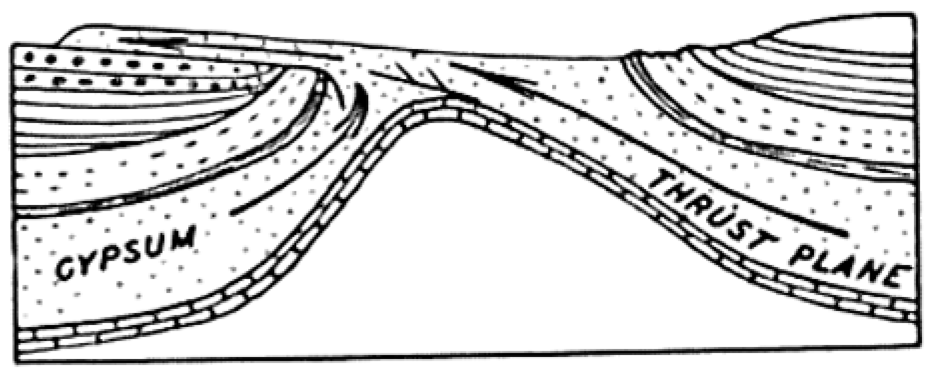


FIG. 85.—AN EROSION THRUST IN WHICH THE MOVING SHEET CONSISTS OF GYPSUM EXPRESSED AT THE SURFACE ALONG THE ERODED CREST OF AN ANTICLINE

(After Busk, *Earth Flexures*)

the gypsum beds is thin[1] (Fig. 85). Erosion of the advancing face of surface thrusts produces coarse detritus, which is deposited in front of the moving thrust sheets and is over-ridden by them, becoming incorporated along the fault plane.

[1] Busk, H. G., *Earth Flexures*: Cambridge, 1929, pp. 86–95.

Goldschmidt has shown that such conditions existed during the growth of the Caledonian thrust nappes in Norway.[1]

Curved Fault Planes.—Fault planes may be curved, either initially or because of later deformation. Initial curvature of shearing planes may result from heterogeneity of the deformed material, or from variation in the orientation of the strain ellipsoid from place to place in homogeneous materials subjected to heterogeneous strain. The loading of one part of a thrust plane by piled up overthrust sheets, or the upward push of puckered incompetent beds beneath the fault, may cause curvature. In incompetent beds, the angle of shear may be different from that in more competent strata, resulting in the deviation of fault planes as they cross from one bed to the next, as noted by Chamberlin and Miller in experimental studies.[2] Curvature of great low angle thrust planes[3] is the rule rather than the exception, and appears to be largely initial curvature. Low angle thrusts folded and faulted by later tectonic movements have, however, been recognized in the southern Appalachians and in the Rockies.[4]

4. *MINOR STRUCTURES ASSOCIATED WITH FAULTS*

Flexures and Folds.—When two crustal blocks begin to move relatively to each other, faulting does not commence immediately. The rocks are first elastically deformed, then a certain amount of plastic yielding occurs, and finally they are sheared through (Fig. 86). During the stages preceding faulting the

[1] Goldschmidt, V., 'Om hoifjeldskvartsen I og II': *Norsk. Geol. Tiddsk.*, Vol. 4, Pt. 1, 1916, pp. 44–6, 49–53. Reviewed in *Geol. Mag.*, Vol. 4, 1917, pp. 130–2.

[2] Chamberlin, R. T., and W. Z. Miller, 'Low-Angle Faulting': *Journ. Geol.*, Vol. 26, 1918, pp. 1–44.

[3] See the literature cited for the Alps, the Scottish Highlands, and the Rockies, on pp. 55–6.

[4] Keith, A., 'Geol. Atlas U.S., Folio No. 151': *U.S. Geol. Surv.*, 1907. Richards, R. W., and G. R. Mansfield, 'The Bannock Overthrust; a Major Fault in South-Eastern Idaho and North-Eastern Utah': *Journ. Geol.*, Vol. 20, 1912, pp. 681–709.

beds are often flexed, and the fault ultimately cuts through the flexure.[1] The beds on either side of the fault plane thus appear as if they have been dragged back by frictional resistance to the movement along the fault, and the flexing is usually referred to as the 'effect of drag'. Only in exceptional circumstances, however, will frictional drag be more important than initial flexing. Such flexing affords a ready means of determining the direction of relative displacement of the blocks (see Fig. 86) and, where developed on a large scale, the flexures are mappable

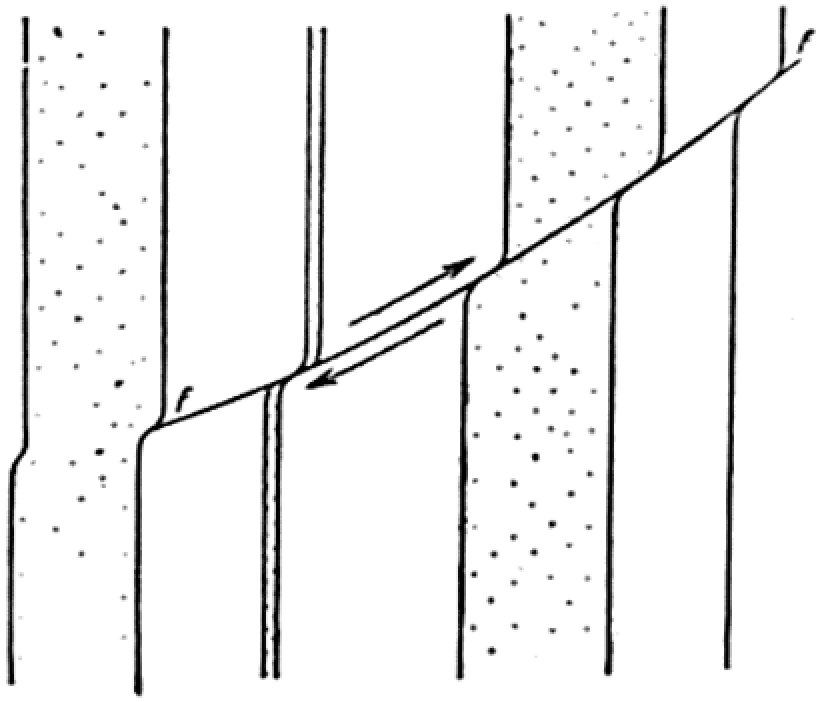


FIG. 86.—A SHEAR THRUST *f–f* CUTTING ACROSS VERTICAL SANDSTONES AND MUDSTONES

On the left the beds are flexed but not faulted, and the fault develops to the right, where the amount of slip progressively increases.

folds. These are best known in the passage of normal faults into monoclines, but the termination of thrust faults may also occur in a fold, generally an isolated flexure.

Close folding associated with low angle thrusts, as in the Belgian coal basins, is regarded as an effect of local compression or confinement due to the curvature of fault planes or to some resistance to the advance of a thrust nappe.[2]

Frictional Effects.—In brittle rocks, noticeable flexing does not occur, and the fault zone is usually brecciated. Considerable movement along a fault may result in extreme granulation

[1] Nádai, A., *Plasticity*: New York, 1931, pp. 300–5.

[2] Kaisin, F., 'Le style tectonique et la genèse mécanique de l'Ardenne': *Bull. Soc. Belge de Géol.*, Vol. 45, 1935, pp. 191–205.

and shearing of the rocks, with the production of *mylonite*. If the finely ground rock is clayey it is called *gouge*; this is frequently polished and striated by the fault movements.

The hanging and foot walls of a fault may be polished, striated, or grooved by rubbing together during the movements. The fine striations are termed *slickensides*, and the larger groovings, *mullion structure*. Slickensides often show a stepped arrangement, the striae terminating abruptly at the edges of the steps, and then continuing again at a lower level. The steps on one wall face the direction towards which the other wall

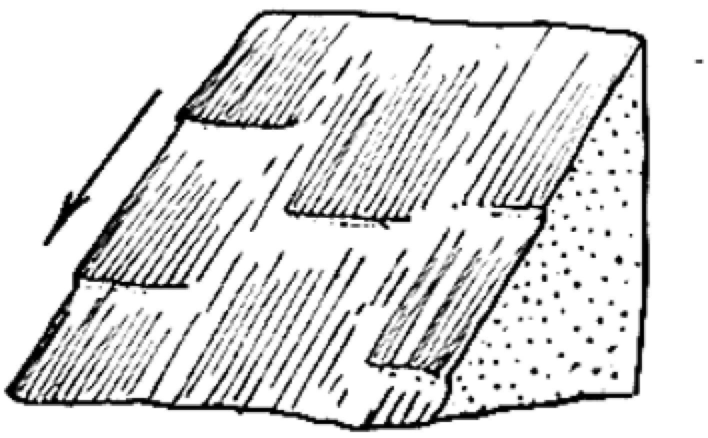


FIG. 87.—SLICKENSIDES ON AN OUTCROP OF SANDSTONE

The steps in the slickensides show that the relative displacement of the hanging wall, which has been removed, was downwards, in the direction of the arrow. Approximately natural size.

moved (see Fig. 87), but if movements took place at different times, the stepped slickensides may indicate only the direction of the last displacement.

Shearing Effects.—Shear and tension joints, subsidiary faults, and fracture cleavage are commonly developed in the rocks adjacent to major faults, and analogous structures have been reproduced in experiments with clay, carried out by Riedel and E. Cloos.[1] In these experiments, two boards are placed on a table, side by side and touching one another, and covered with a cake of clay. One board is then pushed relatively to the other, so that shearing stress is transmitted to the clay above

[1] Riedel, W., 'Zur Mechanik geologischer Brucherscheinungen': *Cbl. f. Min., &c.*, Abt. B, 1929, pp. 354–68. Cloos, E., 'Feather Joints as Indicators of the Direction of Movements on Faults, Thrusts, Joints, and Magmatic Contacts': *Proc. Nat. Acad. Sci. America*, Vol. 18, 1932, pp. 387–95 (with bibliography).

the line of contact of the boards (see Fig. 21, p. 39, and Fig. 88). If the plasticity of the clay is reduced by spraying a film of water on to its upper surface, tension fractures are found to form in the clay above the contact of the two boards. When they first appear, these fractures make an angle of 45°–47°

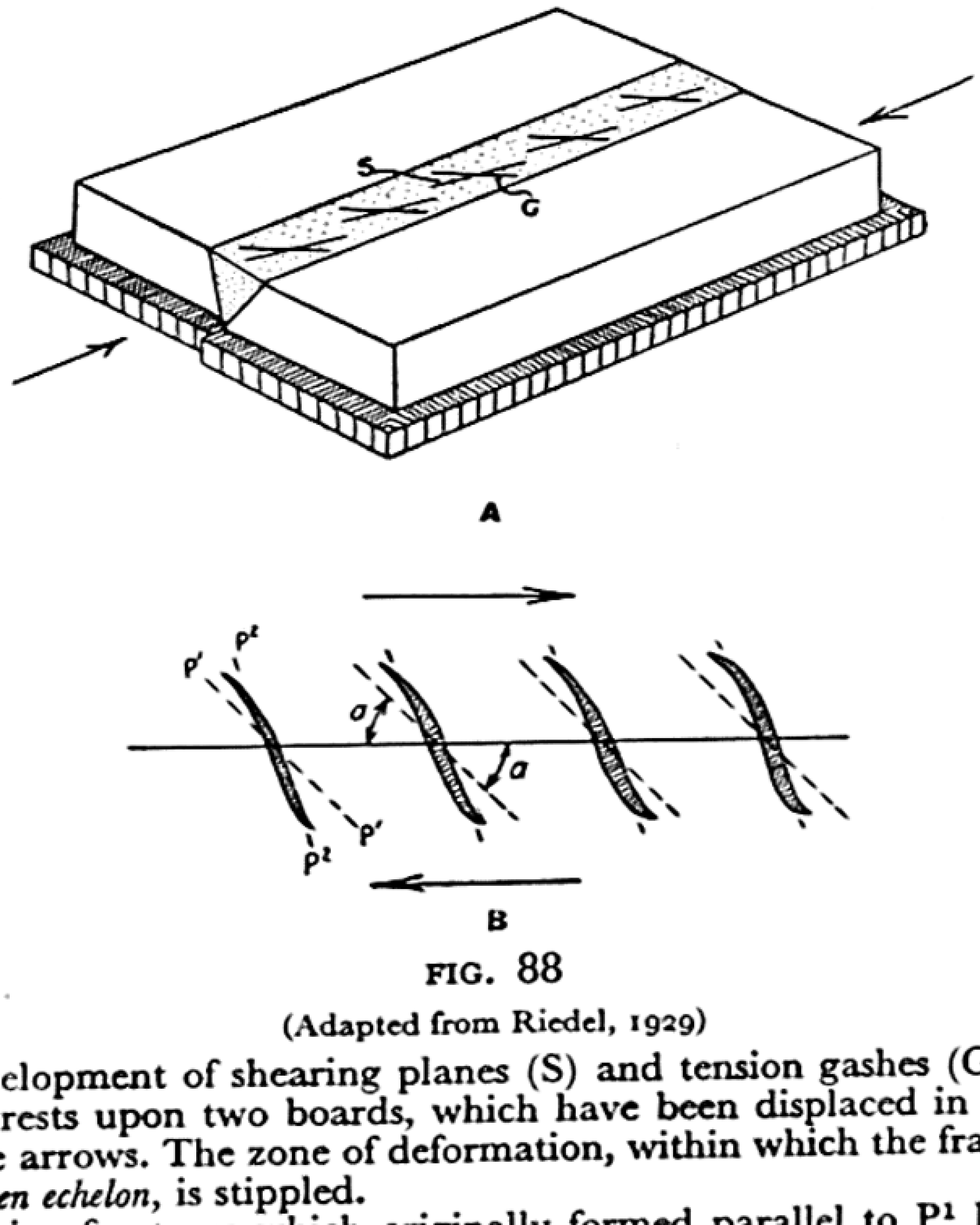


FIG. 88

(Adapted from Riedel, 1929)

A. Development of shearing planes (S) and tension gashes (G) in clay. The clay rests upon two boards, which have been displaced in the direction of the arrows. The zone of deformation, within which the fractures are arranged *en echelon*, is stippled.

B. Tension fractures which originally formed parallel to P^1 have been rotated by continuance of the displacement of the clay in the direction of the arrows, until they have become open gashes parallel to P^2. The acute angles *a* between the gashes and the line along which the displacement of the blocks occurred points in the direction in which the blocks moved.

with the direction along which the relative displacement occurred, but if the experiment is continued, they open out into gashes, and finally may make an angle of as much as 60° with the direction of relative displacement. This is brought about by the rotation of those portions of the clay lying between the first

formed fractures (see Fig. 88, B). By reference to patterns impressed on the upper surface of the clay before the experiment, it can be seen that the tension fractures first form at right angles to the direction of greatest elongation in the deformed zone of clay. If the clay is not flooded with water, but allowed to remain plastic, then shearing planes form in the deformed zone. One dominant set generally appears, making an angle of 12°–17° with the direction of relative displacement of the boards. The complementary set is, however, sometimes feebly developed.

In these experiments it is found that the acute angles enclosed between the tension gashes or the shearing planes, and the direction of relative displacement of the blocks of clay, point in the direction in which the blocks moved. It should be noted that the clay does not shear through along a single surface parallel with the direction of relative displacement in the boards, but that tension gashes and shearing planes are developed *en echelon* in a zone lying between the blocks. A similar arrangement of faults *en echelon* is commonly found in the field, serving to indicate the direction of relative movement of the major crustal blocks on either side of a faulted zone.

Shear and tension joints developed in the zone of deformation between crustal blocks that have moved relatively to each other, e.g. along a fault, are known as *feather joints* (*Federklufte*). The name is given because of the resemblance of the joints and the fault, as seen in cross-section, to the barbs and shaft of a feather. More specifically we may refer to feather joints as *pinnate shear joints* on the one hand, and *pinnate tension joints* (or *gashes*) on the other.

Both field and experimental evidence indicate that feather joints may be used to determine the direction of relative displacement along faults. The rule is that the acute angle enclosed between tension joints and the fault plane points in the direction of movement of the block in which the joints occur (see Fig. 88, B).[1]

[1] Cloos, E., 'Feather Joints as Indicators of the Direction of Movements on Faults, Thrusts, Joints, and Magmatic Contacts': *Proc. Nat. Acad. Sci. America*, Vol. 18, 1932, pp. 387–95.

The marginal thrusts along intrusive contacts (see pp. 148–50) have been interpreted in a similar way, and analogous pinnate structures also occur in the marginal zones of glaciers. In each of these examples we have to deal with masses, of which at least one is plastic and between which there is some degree of cohesion, that have moved relatively to each other like fault blocks.[1]

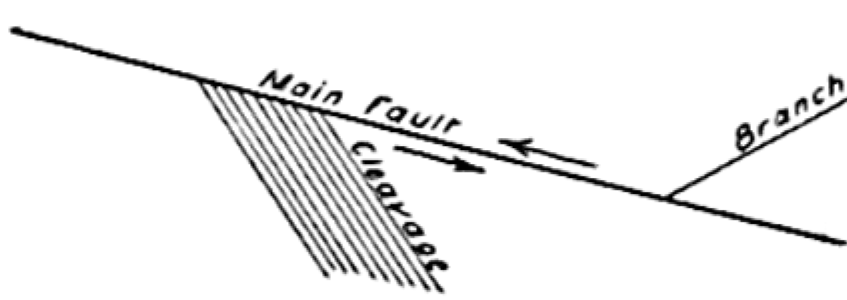


FIG. 89 (A).—FRACTURE CLEAVAGE FORMED IN THE ROCKS ADJACENT TO A FAULT PLANE

(After Sheldon, 1928)

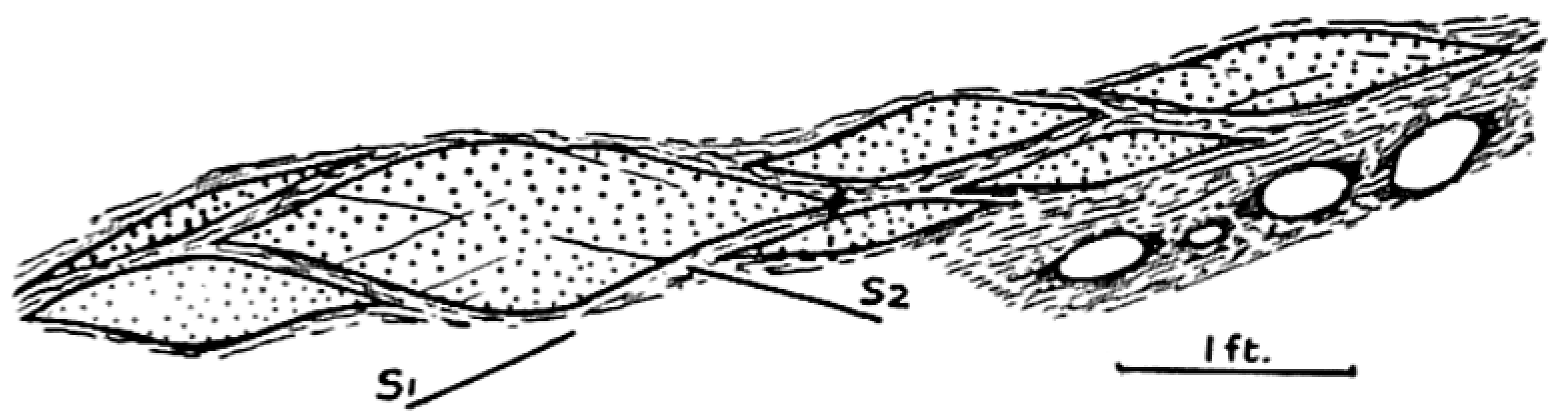


FIG. 89 (B).—PSEUDO-STRETCHED PEBBLES

Shearing of a sandstone bed along directions S1 and S2 forms lenses resembling stretched pebbles. True pebbles (unstretched) in adjacent conglomerate show augen structure due to flow in the matrix.

If pinnate shearing planes along which noticeable displacement has occurred are closely spaced, they constitute cleavage. There is, however, a lack of agreement between the observed relationship of cleavage to faulting, and the relationship that might have been expected by analogy with the above experiments. Sheldon[2] has described an example in which the acute

[1] Cloos, H., 'Zur Mechanik der Randzone von Gletschern, Schollen und Plutonen': *Geol. Rundschau*, 1929, Hft. 1, p. 66, also *Einführung in die Geologie*: Berlin, 1936, pp. 235–6.

[2] Sheldon, P., 'Note on the Angle of Fracture Cleavage': *Journ. Geol.*, Vol. 36, 1928, pp. 171–5.

angle between the cleavage and the fault plane points in the direction opposite to the movement of the block (Fig. 89), and the majority of examples determined in the field agrees with this relationship.

In cases where the fault plane is parallel to the cleavage in the adjacent rocks,[1] it may be that the fault and the cleavage planes are all shearing surfaces induced by the same stress, or that the cleavage planes afforded a predetermined direction of ready shear, which was made use of by later faulting. A *sheeting structure* parallel to the fault plane and confined to the zone of deformation may, however, be a direct result of the faulting, especially in incompetent rocks.

Shear Zones.—Especially in Pre-Cambrian rocks, fault movements may affect zones of considerable width, in which the rocks are crushed, sheared, or even rendered schistose. Such belts are termed *shear* or *crush zones*. Hard rocks involved in the movements are brecciated, often forming crush conglomerate by rounding of the fragments. Where the fracturing of hard layers takes place on two sets of shear planes, the fragments are lenticular, and, if set in finer-grained rocks that flow around them, resemble stretched pebbles. Their origin as disrupted strata may, however, be demonstrated by the alignment of 'pebbles' of similar lithology, representing original strata. (Fig. 89, B).

Faulting and Igneous Activity.—Faulting is closely bound up with igneous activity in all its forms, and a variety of fault types, classified according to their relationships with igneous masses, is recognized. For example, circumferential and radial faults may be connected with volcanic centres or with bysmaliths; ring faults may be occupied by ring dykes, and so on. Some account of these topics is given in Chapter VI.

[1] Dale, T. N., 'Structural Details in the Green Mountain Region, and in Eastern New York': *16th Ann. Rept. U.S. Geol. Surv.*, 1896, Pt. 1, pp. 543–70.

Chapter VI

STRUCTURES OF IGNEOUS ROCKS

LARGELY as a result of the work of the late Hans Cloos and his collaborators, rapid progress has been made of recent years in the study of the megascopic structures of igneous rocks. The advances made have resulted chiefly from the application of the concept that intrusive bodies may be considered from two points of view. They may be considered as fluids which move under stress in chambers within the crust, and as active plungers of plastic material that penetrate the surrounding rocks, as a result of the continued application of the forces that originally caused the injection of the fluid magma. The second concept has proved particularly fruitful in the interpretation of the fracture systems developed in large intrusions. From a consideration of the arrangement of the structures formed during the fluid and plastic stages, much information can be obtained concerning the form of an intrusion and its mechanics of emplacement. The study of these phenomena is included under the term *Granittektonik*, for which *intrusion tectonics* may be suggested as a suitable though rather more comprehensive equivalent in English.[1] Methods of research

[1] The following are works of a general nature dealing with intrusion tectonics: Cloos, H., *Der Mechanismus tiefvulkanischer Vorgänge*: Braunschweig, 1921, Sammlung Vieweg; 'Das Batholithenproblem': *Fortschr. d. Geol. u. Pal.*, Hft. 1, 1923, p. 80; *Einführung in die tektonische Behandlung magmatischer Erscheinungen*: Pt. 1, Das Riesengebirge: Berlin, 1925 (with bibliography to 1925); 'Einige Versuche zur Granittektonik': *Neues Jahrb. f. Min., &c.*, Beil. Bd. 64, Abt. A, 1931, pp. 829–36. Grout, F. F., 'Scale Models of Structures Related to Batholiths': *Amer. Journ. Sci.*, Vol. 243A, 1945, pp. 260–84. Balk, R., 'Primary Structures of Granite Massives': *Bull. Geol. Soc. Amer.*, Vol. 36, 1925, pp. 679–96; 'Structural Behaviour of Igneous Rocks': *Mem. Geol. Soc.*

used in the study of intrusion tectonics have also been applied to extrusive rocks, and these will be dealt with as occasion arises.

1. *STRUCTURES DUE TO FLOW*

When a magma advances from one point to another as a liquid stream, as in the extrusion of lavas and the injection of intrusive bodies along planes of weakness in the crust, the movement takes place mainly by stream-line flow;[1] that is, every particle passing a given point takes the same path or stream-line. Variation in the viscosity of the magma from point to point, due to temperature and composition differences,[2] and variation in the stress, or in the frictional resistance to flow developed at interfaces, all combine to cause differences in the rate of movement of particles following adjacent stream-lines. If inclusions which have dimensional axes of unequal length are present in the liquid, they will be rotated as a result of the differential stream-line flow until they lie with their longest axes parallel to the stream-lines, thus presenting a minimum of resistance to the fluid motion. The parallel orientation of needle-shaped inclusions constitutes what is termed *linear flow structure* in an igneous rock, and this orientation, by revealing to the eye the *flow lines* formerly present in the fluid magma, allows the direction of flow to be determined at each point of observation (see, however, p. 138).

On the other hand, parallelism of the flat surfaces of tabular or platy inclusions such as phenocrysts or xenoliths, or of *schlieren*, constitutes *platy flow structure. Schlieren* are layers that contain the same minerals as the average of the whole rock mass, but in different proportions.[3] Concentrations of dark minerals give melanocratic *schlieren*, and of light-coloured minerals, leucocratic *schlieren*, so that a rock in which these are

Amer., No. 5, 1937. The latter is the most complete account of the subject in English.

[1] Bond, W. N., *An Introduction to Fluid Motion*: London, 1925, pp. 11–12.

[2] Balk, R., 'Viscosity Problems in Igneous Rocks': *Journ. Rheology*, Vol. 3, 1932, pp. 461–78.

[3] This statement may not always be strictly applicable if the exact composition of mixed crystals is considered.

well developed appears distinctly banded. The term *flow layer* has been suggested as the English equivalent of *schliere*,[1] and it may be useful to restrict the use of the term *foliation*, when dealing with igneous rocks, to platy flow structures revealed by flow layers and by tabular inclusions. Surfaces of foliation are developed parallel to any contacts which exert friction on a moving magma. In a lava they are therefore formed parallel to the base of the flow, and in an intrusion, with rare exceptions,

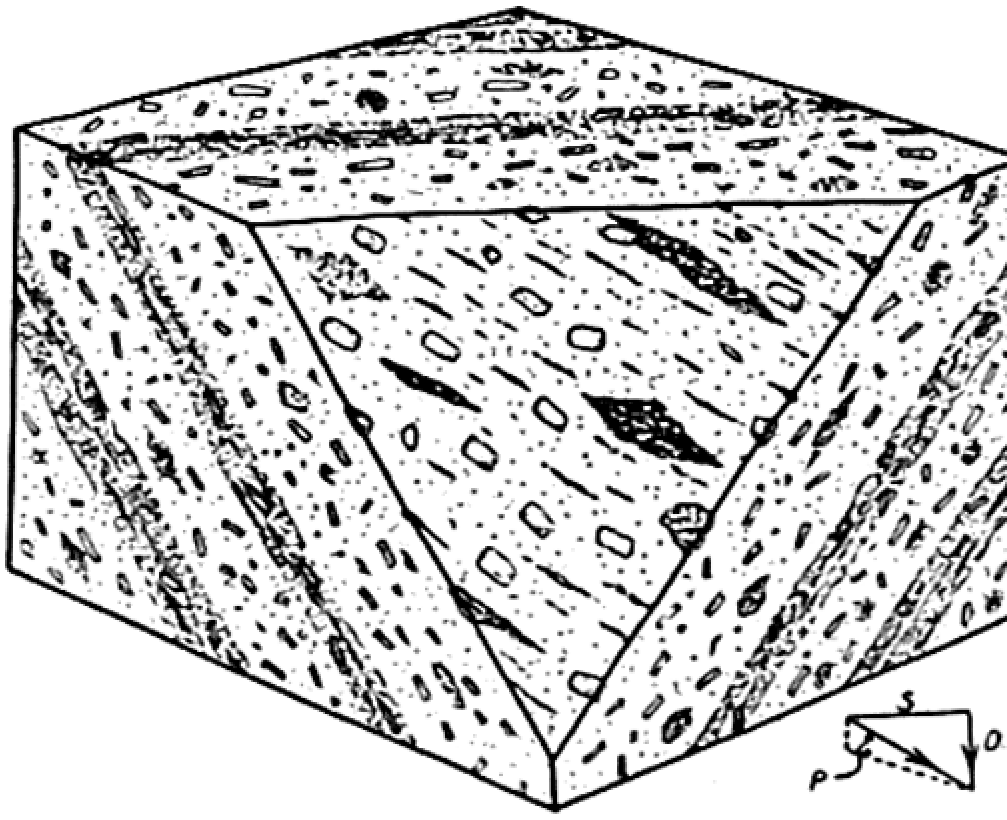


FIG. 90.—BLOCK OF PORPHYRITIC GRANITE SHOWING PLATY AND LINEAR FLOW STRUCTURES

Melanocratic flow layers dip at a high angle towards the observer. Foliation in the granite itself is shown by the parallel orientation of platy xenoliths and tabular phenocrysts. Flow lines, pitching at an angle P which is less than the dip of the foliation planes, are revealed by the parallel orientation of linear inclusions. S is the strike of the foliation, D its direction of inclination.

parallel to the contacts. Both flow lines and foliation may be present in the same rock, or one or the other may be absent. If both are developed, the flow lines lie in the planes of foliation, but may make any angle with the dip of these planes (Fig. 90).

It has been pointed out by H. Cloos that the observed parallel orientation of needle-shaped inclusions in an igneous

[1] Balk, R., 'Structural Behaviour of Igneous Rocks': *Mem. Geol. Soc. Amer.*, No. 5, 1937, p. 15.

rock does not necessarily imply the former presence of stream-lines in the magma, parallel to the direction of alignment of the inclusions. As is well shown by the foam streaks on mountain torrents,[1] objects lying on the surface of a broad stream tend to arrange themselves in arcs whose convex sides face the direction in which the stream is moving. Thus, in the centre of the stream the flow structures are at right angles to the general direction of flow, while near the banks they are parallel or nearly so, to this direction, owing to the frictional drag. If a line of particles that at first stretched directly across the stream from bank to bank is considered, it will be found that this line becomes curved as it moves downstream, and at the same time it is stretched along its length to accommodate itself to the arching (see Fig. 91). The effects of this stretching are of considerable importance in the interpretation of the fracture systems in intrusions (see pp. 144–50). The *flow wrinkles* on the upper surfaces of small lava tongues afford a good example of the arching of flow structures, and they serve to indicate the local direction of flow in ancient lavas. These wrinkles have, moreover, a further practical value in structural field work, owing to the fact that their crests are rounded and the troughs between them V-shaped. This aids in the recognition of the upper surfaces of lava flows in disturbed

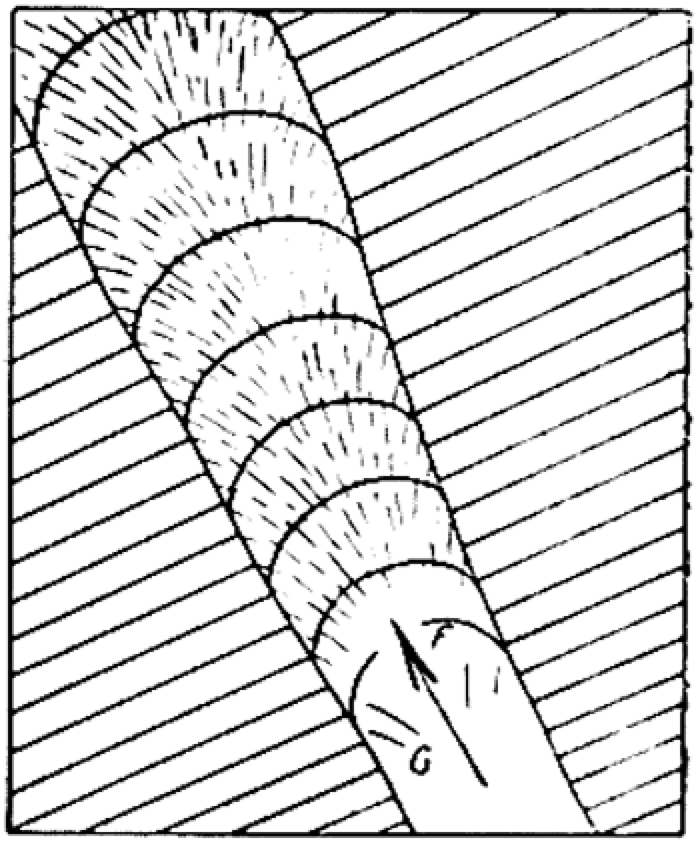


FIG. 91.—ARCHED FLOW STRUCTURES (F), WITH TENSION GASHES (G) AT RIGHT ANGLES TO THEM, IN A STREAM FLOWING IN THE DIRECTION OF THE ARROW

The figure may be used to represent crevasses (G) and transverse ridges (F) on a valley glacier; 'pressure ridges' (F) on the surface of a lava flow; curved flow lines (F) in a dyke; *Q*-joints (G) and a dome of flow lines (F) in a large intrusion; flow wrinkles (F) on a small tongue of basalt.

[1] Balk, R., *op. cit.*, 1937, Plate I.

rocks,[1] but the test is not always reliable, for similar flow wrinkles sometimes form on the lower surfaces of flows that rest on dry scoriae and ashes (Fig. B, Pl. II).

Especially in the study of Pre-Cambrian basic lava flows, much use has been made of pillow structure in determining facings. In North America the method has been applied chiefly to Keewatin volcanics, and in western Australia to lavas believed to be of similar age at Kalgoorlie, Coolgardie and elsewhere. The method depends on the assumption that the upper surfaces of individual 'pillows' are different from lower. In general, it is assumed that the former are rounded while the latter are cuspate, often with a 'tail' that fits between the rounded tops of lower pillows. Experience shows that although particular examples may not yield definite information, the method generally yields consistent results.[2]

It is found that the foliation surfaces and flow lines in large intrusions commonly form a dome or arch. The term *dome* is used if the flow structures extend over the whole massif, but if they are absent from the interior, and exist only around the borders, the arrangement is termed an *arch*. These structural patterns are probably caused by the upward movement of an igneous mass as a whole under the influence of pressure from below, the domes or arches being formed in an analogous manner to the foam streaks and flow wrinkles described above. In dealing with large intrusions, it appears to be preferable to consider the flow lines as indicating the direction of stretching (the long axis of a form ellipsoid), rather than simply as stream-lines (see Fig. 92). In the border zones of intrusions, foliation surfaces are often especially well developed, and the rock may be gneissic. Flow lines, where observed in such gneissic border zones, pitch at the steepest possible angle, that is at the angle of dip of the foliation. In the central parts of massifs flow

[1] Tanton, T. L., 'Determination of Age-Relationships in Folded Strata': *Geol. Mag.*, Vol. 67, 1930, pp. 73–6. See also Butler, B. S., and W. S. Burbank, 'The Copper Deposits of Michigan': Prof. Paper No. 144, *U.S. Geol. Surv.*, 1929, for a good account of the features of flow tops.

[2] See Wilson, M. E., 'Structural Features of the Keewatin volcanic rocks of Western Quebec': *Bull. Geol. Soc. Amer.*, Vol. 53, 1942, pp. 54–69 (with bibliography).

structures usually fade out, but they may persist throughout small intrusions, and in these a dome or arch is generally absent.[1] The absence of a structural dome in large intrusions signifies in most cases that erosion has removed the roof rock and those parts of the intrusion adjacent to it.

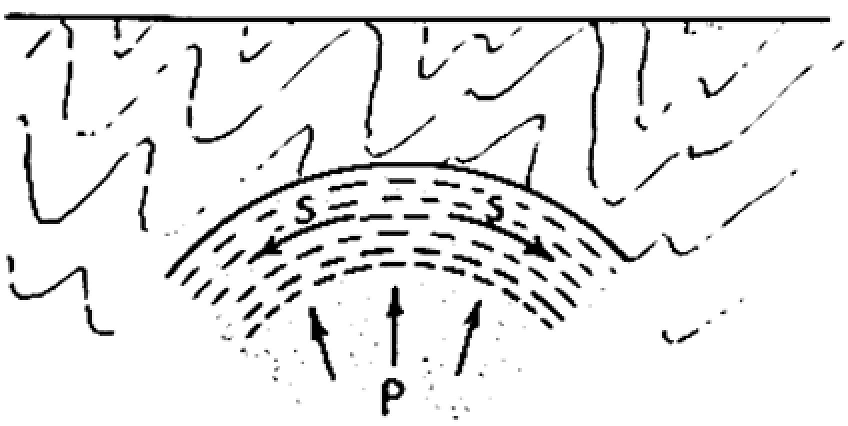


FIG. 92.—DIAGRAMMATIC CROSS-SECTION OF A BATHOLITH, SHOWING THE DOME OF FLOW STRUCTURES CAUSED BY THE PRESSURE P, WITH STRETCHING IN THE DIRECTION OF THE ARROWS

Examples of the mapping and interpretation of flow structures will be found in the works listed below[2] (see also Figs. 93 and 94).

[1] Grout, F. F., and R. Balk, 'Structural Study of the Snowbank Stock': *Bull. Geol. Soc. Amer.*, Vol. 45, 1934, pp. 621–36.

[2] VOLCANIC ROCKS—Cloos, H., and E. Cloos, 'Die Quellkuppe des Drachenfels am Rhein': *Zeitschr. f. Vulkanologie*, Vol. 11, 1927, pp. 33–40. Allen, J. E., 'Structures in the Dacitic Flows at Crater Lake, Oregon': *Journ. Geol.*, Vol. 44, 1936, pp. 737–44. INTRUSIONS—Balk, R., 'A Contribution to the Structural Relations of the Granitic Intrusions of Bethel, Barre, and Woodbury, Vermont': *Biennial Rept., Vermont State Geologist*, Vol. 15, 1926; 'Die Primäre Struktur des Noritsmassivs von Peelskill am Hudson': *Neues Jahrb. f. Min., &c.*, Beil. Bd. 57, Abt. B, 1927, pp. 249–303; 'Structural Geology of the Adirondack Anorthosite': *Tscherm. Min. Pet. Mitt.*, Bd. 41, 1931, pp. 309–434; 'Inclusions and Foliation in the Harney Peak Granite, Black Hills, South Dakota': *Journ. Geol.*, Vol. 39, 1931, pp. 736–48. Cloos, E., 'Structural Survey of the Granodiorite South of Mariposa, California': *Amer. Journ. Sci.*, Vol. 23, 1932, pp. 289–304; 'Der Sierra-Nevada-Pluton in Californien': *Neues Jahrb. f. Min., &c.*, Beil. Bd. 76, Abt. B, 1936, pp. 355–450. Mayo, E. B., 'Some Intrusions and their Wall Rocks in the Sierra Nevada': *Journ. Geol.*, Vol. 43, 1935, pp. 673–89. Osman, C. W., 'The Granites of the Scilly Isles and their Relation to the Dartmoor Granites': *Quart. Journ. Geol. Soc.*, Vol. 84, 1926, pp. 258–92.

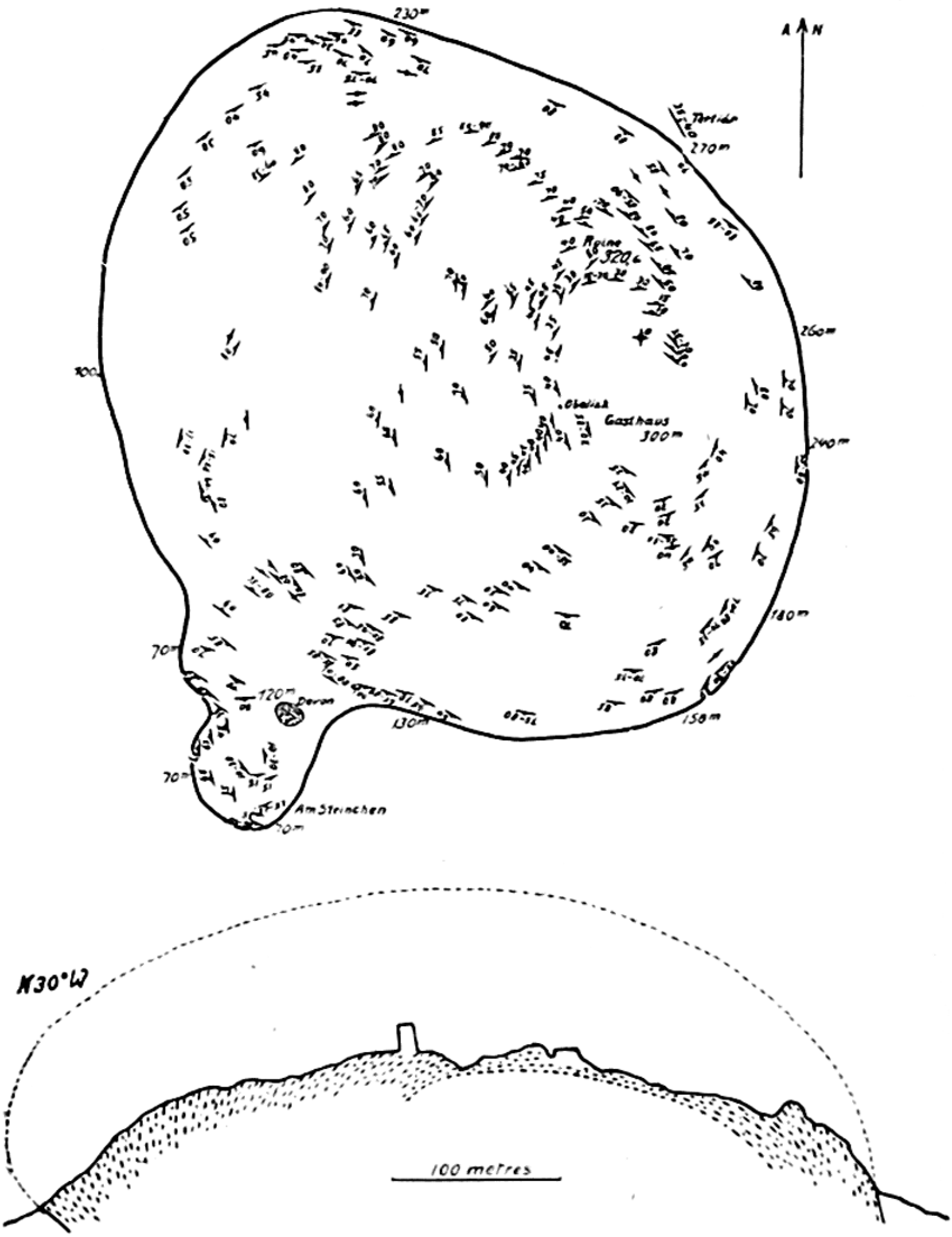


FIG. 93.—DOME OF PLATY FLOW STRUCTURES, SHOWN BY THE PARALLEL ORIENTATION OF TABULAR FELSPAR PHENOCRYSTS IN THE DRACHENFELS TRACHYTE

(After H. and E. Cloos, 1927)

The section shows the reconstructed form of the trachytic body, which is a swelling at the upper end of a volcanic neck, beneath a cover of tuffs.

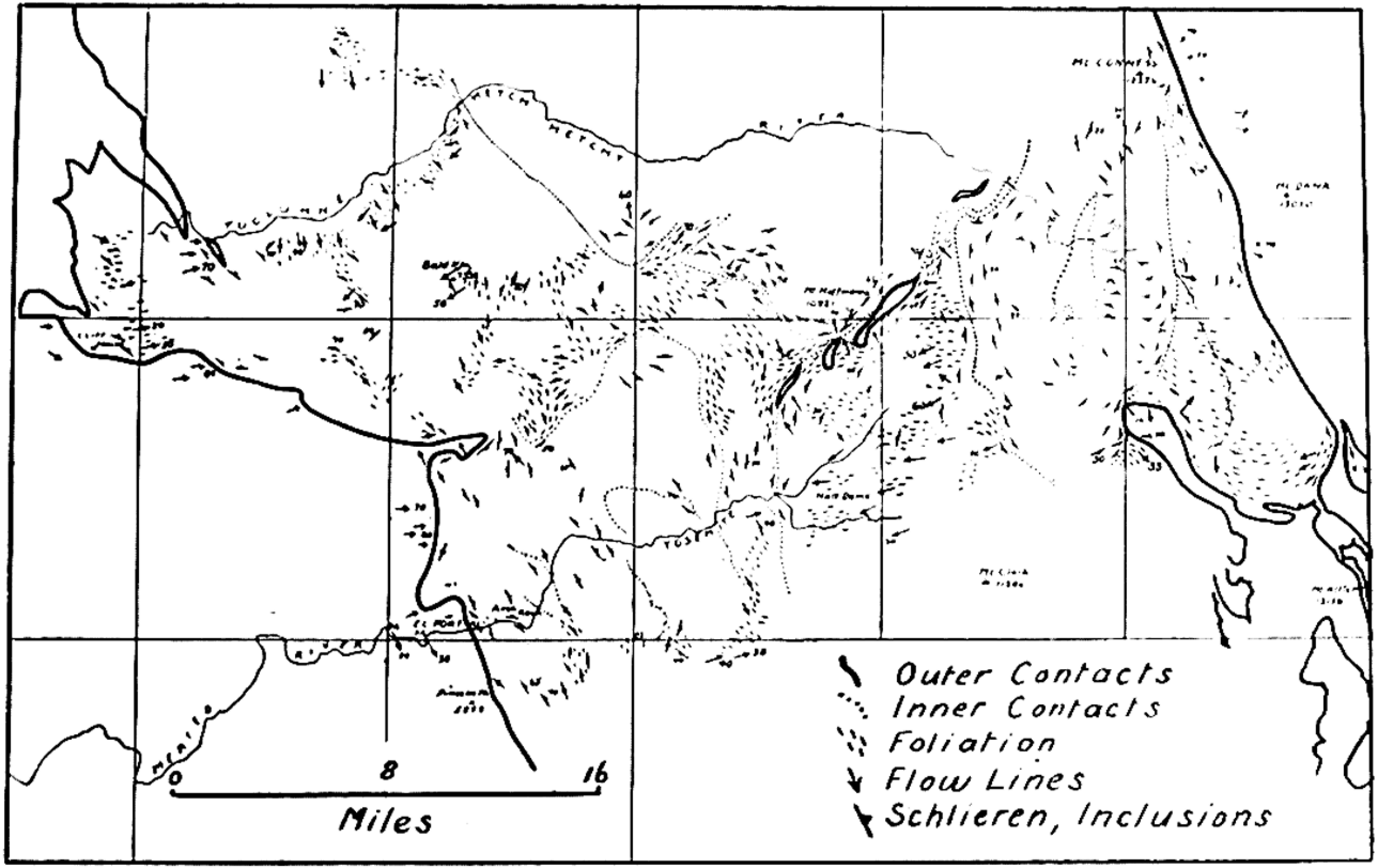


FIG. 94.—STRUCTURAL MAP OF THE SIERRA NEVADA BATHOLITH, CALIFORNIA

(After E. Cloos, 1936)

2. *STRUCTURES DUE TO FRACTURE*

Joints and faults can arise in igneous rocks only when these have attained the solid state, or when a crystal mesh has formed, and they may then be caused by continued application of the same forces that led to the uprise of the magma in the crust, by deformation at a later date by entirely unrelated forces, or by stresses set up as a result of contraction during cooling. Jointing due to the latter cause is particularly well exhibited by volcanic rocks, especially basalts, in the form of *columnar jointing*. The rock is divided into columns, generally hexagonal in cross-section, but sometimes four, five, or seven-sided, the long axes of the columns being arranged at right angles to cooling surfaces.[1] Similar columnar jointing is also shown in some dykes, the columns lying at right angles to the dyke walls. In plutonic intrusions, cooling joints have not been recognized with certainty, although the characteristic vertical joints in granites have been ascribed to this cause by earlier workers.[2] H. Cloos and his co-workers believe that most of the joints and minor faults in granite massifs arise during the later cooling stages of the intrusions, owing to the persistence of stresses which caused the magma to rise in the crust. Cloos was led to this conclusion by the discovery that many of the structures of intrusions in Europe and North America appear to have a definite geometrical relationship to the form of the intrusive bodies, and later work has substantiated most of his ideas, although they have met with objection from some quarters.[3]

[1] James, A. V. G., 'Factors producing Columnar Structure in Lavas, and its occurrence near Melbourne, Victoria': *Journ. Geol.*, Vol. 28, 1920, pp. 458–69 (with bibliography). Hunt, C. B., 'A Suggested Explanation of the Curvature of Columnar Joints in Volcanic Necks': *Amer. Journ. Sci.*, Ser. 5, Vol. 36, 1938, pp. 142–9.

[2] Geikie, J., *Structural and Field Geology*: Edinburgh, 1905, p. 149. It is possible that *S*-joints in granites (see p. 146) may be caused by contraction on cooling—see Cloos, H., 'Über Ausbau und Anwendung der Granittektonischen Methoden': *Abh. d. Preuss. Geol. Landesanst.*, N.F., Vol. 89, 1922, p. 5.

[3] See Sander, B., 'Zur Granittektonik, Mikrotektonik, &c.': *Verh. Geol. Bundesanst. Wien*, 1923; also Cloos's reply in *Cbl. f. Min., &c.*, 1926, Abt. B, pp. 481–92.

The border zones of intrusions are particularly suitable for the investigation of structures, because the rocks there tend to be more heterogeneous than in the central parts of the intrusions, and they thus present clearer optical evidence of small displacements. Also, the border zones become plastic while the central parts are still fluid, and fractures which arise in them are therefore likely to become injected with magmatic differentiates, thus rendering the fractures visible even in homogeneous rocks.

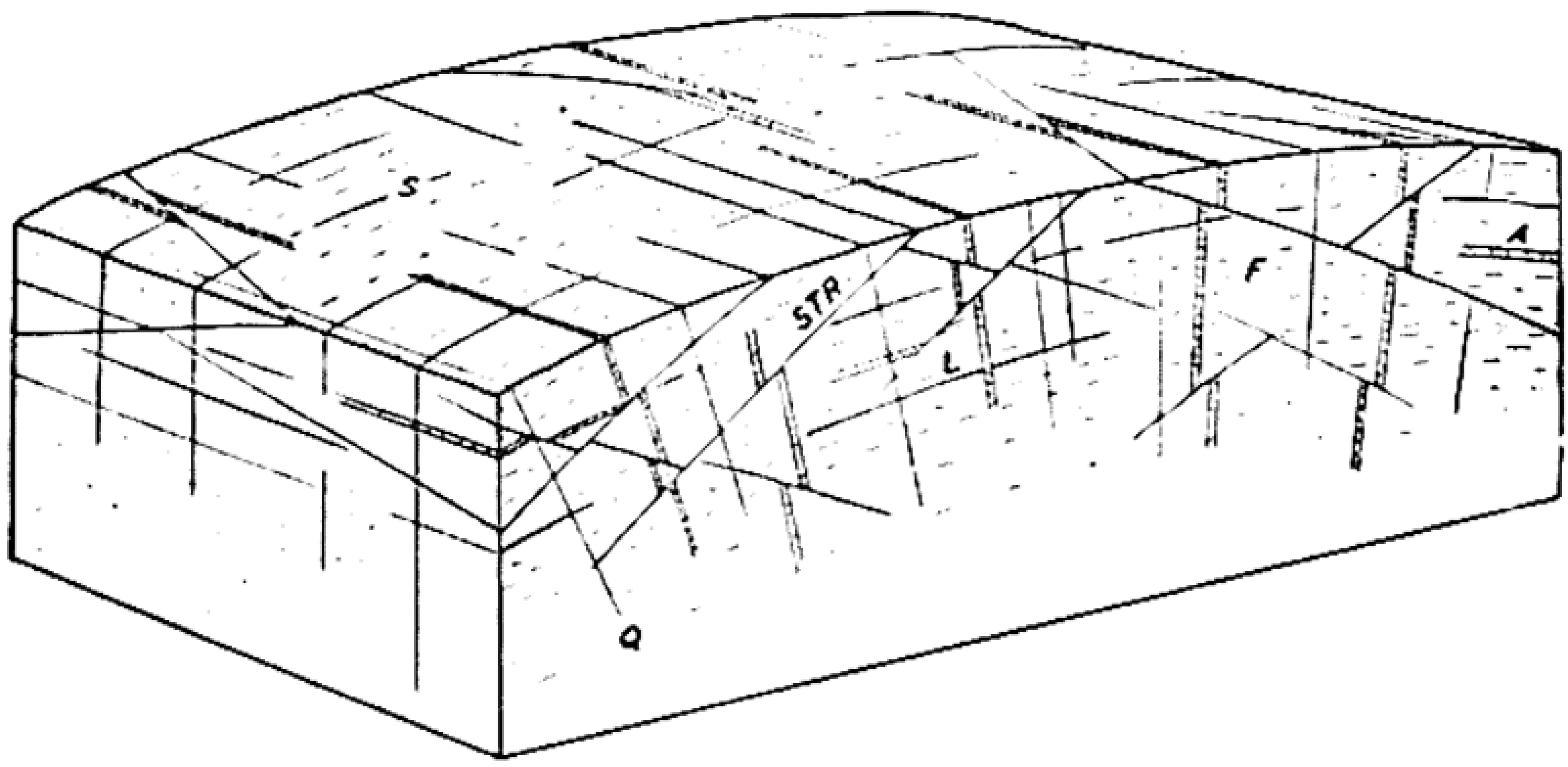


FIG. 95.—BLOCK DIAGRAM SHOWING THE CHIEF TYPES OF JOINTS IN A BATHOLITH

(After H. Cloos, 1923)

Q, cross joints; S, longitudinal joints; L, flat-lying joints; *Str.* planes of stretching; F, linear flow structures; A, aplite dykes.

Joints.—Joints which lie perpendicular to the flow lines are termed *Q*—or *cross joints*[1] (see Fig. 95). These are tension joints, formed as a result of the upward push transmitted to the already solidified rocks adjacent to the contacts, by the liquid or highly plastic magma in the central parts of an intrusion. As we have seen, the flow lines represent the direction of 'stretching' in the marginal zones of an intrusion, and the opening of cross joints as tension gashes permits this stretching

[1] Cloos, H., 'Tektonik und Magma, Bd. 1': *Abh. d. Preuss. Geol. Landesanst.*, N.F., Vol. 89, 1922. Balk, R., 'Structural Behaviour of Igneous Rocks': *Mem. Geol. Soc. Amer.*, No. 5, 1937, pp. 27–34.

to continue during the early stages of solidification. If a dome or arch is developed in the flow lines, the cross joints are arranged like the ribs of a fan, and converge towards the centre of the arch.[1] Joints having the characters of *Q*-joints may be found in rocks in which flow lines are absent, and it is then recommended that they be termed *tension joints.*[2]

S-joints (*Längenklüfte, longitudinal joints*) are steeply dipping joints that strike parallel to the flow lines as projected on to a map, and they are best developed where the flow lines approach the horizontal, which is often the case near the roof of a large intrusion. That the *Q*- and *S*-joints are formed during the transition of the rock mass from the fluid state to the solid, is often indicated by the presence along them of aplites, pegmatites, or other dyke rocks genetically related to the intrusion. The mode of formation of *S*-joints is not clearly understood.

Flat-lying joints (*Lagerklüfte*) in large intrusions are of two types—those formed during the emplacement of the massif (primary flat-lying joints) and those of later origin. The primary joints, according to Balk,[3] may or may not embrace the flow lines, where these are gently inclined, and if these joints are not filled with vein or dyke rocks their origin as primary structures cannot be readily demonstrated. Flat-lying joints of subsequent origin may be caused either by the processes of weathering and erosion, or by horizontal compression acting on the solid massif. A well-developed *sheeting*, consisting of gently curved joints that divide the rock up into flat lenses, lying approximately parallel to the topographic surface at each locality, is generally present in granites. If closely spaced, the sheeting joints constitute *mural jointing*. It is sometimes found that granite showing well-developed sheeting joints is under compression, as is shown by the expansion, at times with explosive violence, of quarried blocks. In such cases the

[1] Cloos, E., 'Der Sierra-Nevada-Pluton in Californien': *Neues Jahrb. f. Min., &c.*, Beil. Bd. 76, Abt. B, 1936, p. 392.

[2] Balk, R., 'Structural Behaviour of Igneous Rocks': *Mem. Geol. Soc. Amer.*, No. 5, 1937, p. 33.

[3] Balk, R., *op. cit.*, pp. 39–42.

sheeting joints may be potential shearing planes formed by horizontal compression under the action of diastrophic forces.[1] Where the sheeting reflects the surface topography of the granitic area, however, it is probably caused by a combination of such factors as the expansion of the felspars and ferro-magnesian minerals on weathering, removal of the load of superincumbent rock by erosion, and seasonal temperature variations affecting the rock near the surface. Sheeting due to these causes becomes less pronounced in depth.

Rift, Grain, and Hardway.—Strict correlation of the terminology of natural fracture planes with that used to describe the directions along which granites are split by quarrymen is not possible, because the terms used for the artificial planes of parting have no fixed and precise significance. Generally, the term *rift* is applied to the direction of most ready parting, and the term *grain* to another direction of ready parting, generally lying approximately at right angles to the rift. The *hardway* is the third direction, along which it is necessary to split the rock in order to obtain a block from the quarry, and it is more difficult than the other two.[2] There is wide divergence in the usage of these terms, however, and it is always necessary to determine the local usage in studying a quarry.[3]

The rift in some granites has been shown to be determined by minute parallel cracks in the quartz crystals,[4] and also, to a certain extent, by parallel strings of bubbles in the quartz. In the Pre-Cambrian granites of Quebec,[5] the rift (as defined above) is inclined in the same direction as the surface slopes of the granitic area, and is thought to have been caused by a

[1] Dale, T. N., 'The Granites of Vermont': *U.S. Geol. Surv.*, Bull. No. 404, 1909, pp. 17–18; 'The Commercial Granites of New England': *U.S. Geol. Surv.*, Bull. No. 738, 1923.

[2] Dale, T. N., *op. cit.*, 1923. Howe, J. A., *The Geology of Building Stones*: London, 1910, pp. 54–8.

[3] For example, Balk defines the *grain* as steep planes along which the ease of splitting is as good as, or better than, the rift.

[4] Dale, T. N., 'The Commercial Granites of New England': *U.S. Geol. Surv.*, Bull. No. 738, 1923. Bell, J. F., 'The Investigation of the Cleavage of Granites': *Econ. Geol.*, Vol. 31, 1936, pp. 272–7.

[5] Osborne, F. F., 'Rift, Grain, and Hardway in some Pre-Cambrian Granites, Quebec': *Econ. Geol.*, Vol. 30, 1935, pp. 540–51.

combination of the effects of temperature changes near the surface, and of the removal of superincumbent granite by erosion. In this instance it therefore has no connexion with the primary structures of the intrusion, and rapidly becomes ill-defined in depth. According to H. Cloos,[1] the rift in granites usually coincides with the direction of the *S*-joints.

Typically the grain is parallel to the planes of foliation. Grain planes therefore show the flow structure in the rock, which, owing to its resemblance to the grain in wood, gives them their name. The hardway may be a random direction determined by the attitude of the rift and grain, and necessary to the quarrymen in order to obtain rectangular blocks, or it may correspond in some cases with the direction of the *Q*-joints, and is then to be regarded as a primary structure. It is found that these *Q*-joints are more constant in direction than the grain in the Quebec granites, and this agrees with Cloos's idea of the origin of *Q*-joints as tension joints formed by warping of the mass as a whole in the solid state.

Faults.—Both thrust and normal faults occur in the border zones of large intrusions. The thrusts, which have been termed *marginal thrusts*, traverse both the intrusion and its wall rocks, and are arranged *en echelon* around the margins of the igneous body. The amount of the slip along each thrust is usually small, of the order of inches rather than of feet, but as they occur in great numbers their total effect is considerable.[2] Movements along marginal thrusts are directed outwards from the intrusive centre, and the thrusts are regarded by E. Cloos as pinnate shearing planes developed as a result of the rise of the plastic intrusion relatively to the adjacent wall rocks. They may thus be interpreted, by analogy, in the same way as the shear joints along fault lines and the shearing planes in clay produced in

[1] Cloos, H., 'Über Ausbau und Anwendung der Granittektonischen Methoden': *Abh. d. Preuss. Geol. Landesanst.*, N.F., Vol. 89, 1922, pp. 2–4.

[2] Cloos, E., 'Feather Joints as Indicators of the Direction of Movements on Faults, Thrusts, and Magmatic Contacts': *Proc. Nat. Acad. Sci. Washington*, Vol. 18, 1932, pp. 387–95; 'Der Sierra-Nevada-Pluton in Californien': *Neues Jahrb. f. Min., &c.*, Beil. Bd. 76, Abt. B, 1936, pp. 355–450. Balk, R., 'Structural Behaviour of Igneous Rocks': *Mem. Geol. Soc. Amer.*, No. 5, 1937, pp. 101–6.

the experiments carried out by Riedel and Cloos (see pp. 39, 132). Minor sets of pinnate shear joints may arise as a result of

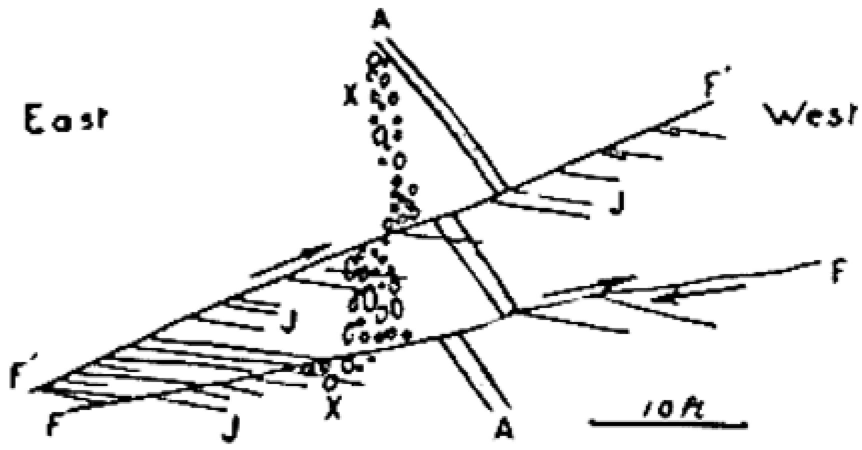


FIG. 96.—MARGINAL THRUSTS (F, F′) TOWARDS THE WESTERN CONTACT OF PART OF THE SIERRA NEVADA BATHOLITH

(After E. Cloos, 1932)

A, aplite; J, feather joints along the thrusts; X, swarm of xenoliths.

the movement along the thrusts themselves (see Fig. 96). At steep contacts, marginal thrusts dip at low angles, but nearer

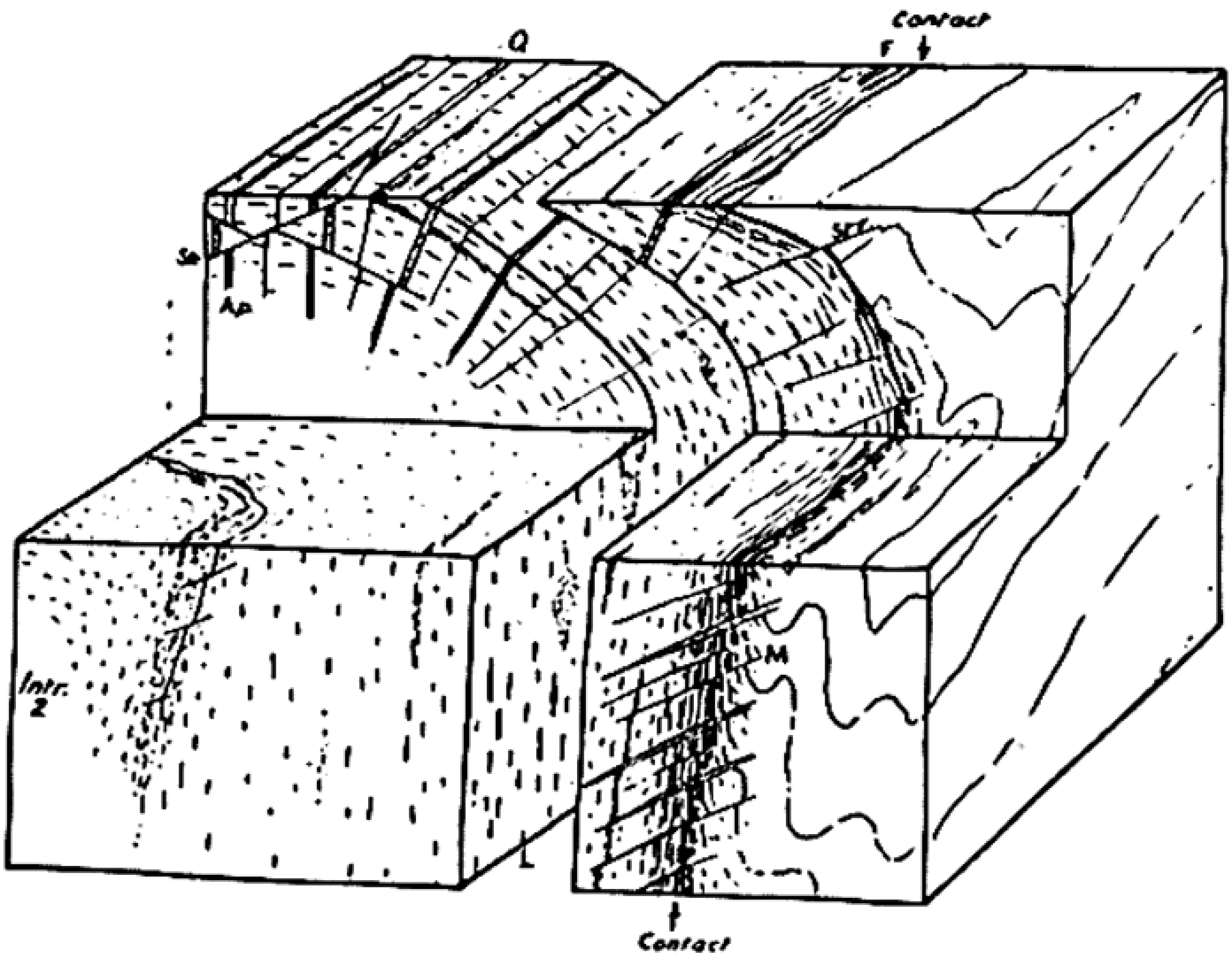


FIG. 97.—BLOCK DIAGRAM OF PART OF A BATHOLITH, WITH A SUBSIDIARY INTRUSION, DISSECTED ALONG A FOLIATION SURFACE

M, marginal thrusts, some with injected aplite; F, flow layers and foliation; L, linear flow structure; Q, cross joints, some with injected aplite; *Str.*, planes of stretching. Schistosity parallel to the granite contact is developed in the wall rocks, and the axes of folds are tilted away from the intrusion.

the roof of an intrusion, where the contact is flat-lying, the thrusts dip at higher angles. If the physical condition of the rocks in contact zones is more favourable to the formation of tension gashes than of shearing planes—that is, if they are relatively brittle—marginal fissures, often filled with dyke rocks, develop. These may be distinguished from cross joints by the fact that the latter lie at right angles to the flow lines, whereas the fissures cut obliquely across them.[1]

Flat-lying normal faults (*Streckfläche, planes of stretching*) are developed as complementary shearing planes which may be regarded as formed by the stretching of the border zones of large intrusions, parallel to the flow lines. The sum of the movements along them constitutes an extension of the mass in an approximately horizontal direction, and they are usually restricted to the upper parts of intrusions. The corresponding vertical extension of the mass is effected by the marginal fissures and thrusts found along steep contacts.

The geometrical relationships of the structures of large intrusions are shown diagrammatically in Figs. 95 and 97.

3. *TECTONIC RELATIONSHIPS OF LARGE INTRUSIVE BODIES*

For structural purposes it is necessary to distinguish between three major realms of intrusive igneous activity[2] (see Fig. 98).

(1) In the lowest levels of the crust—the zone of abyssal intrusions, or ultraplutonic zone—the country rock is highly plastic or even fluid, the mass of superincumbent rocks is very great, and small stress differences are sufficient to cause considerable movements. Ultraplutonic intrusions are characterized by the dominance of structures due to flow, both in the

[1] See Fig. 88. Imagining the zone between the blocks of clay as the intrusive contact, with which the flow lines will be parallel, it will be seen that the tension gashes will cut the flow lines.

[2] See Kober, L., *Der Bau der Erde*: Berlin, 1928, pp. 38–80. Cloos, H., *Einführung in die Geologie*: Berlin, 1936, pp. 75–82. Bucher, W. H., *The Deformation of the Earth's Crust*: Princeton, 1933, pp. 267–302. Evans, J. W., 'Regions of Tension': *Quart. Journ. Geol. Soc.*, Vol. 81, 1925, pp. lxxx–cxxii; 'Regions of Compression': *ibid.*, Vol. 82, 1926, pp. lx–cii.

intrusions and their wall rocks, and fracture phenomena are absent. Injection of magma occurs along zones of weakness, and assimilation, granitization, or actual melting of the country rocks is widespread. These features are shown by many of the Pre-Cambrian rocks exposed in different parts of the world.[1]

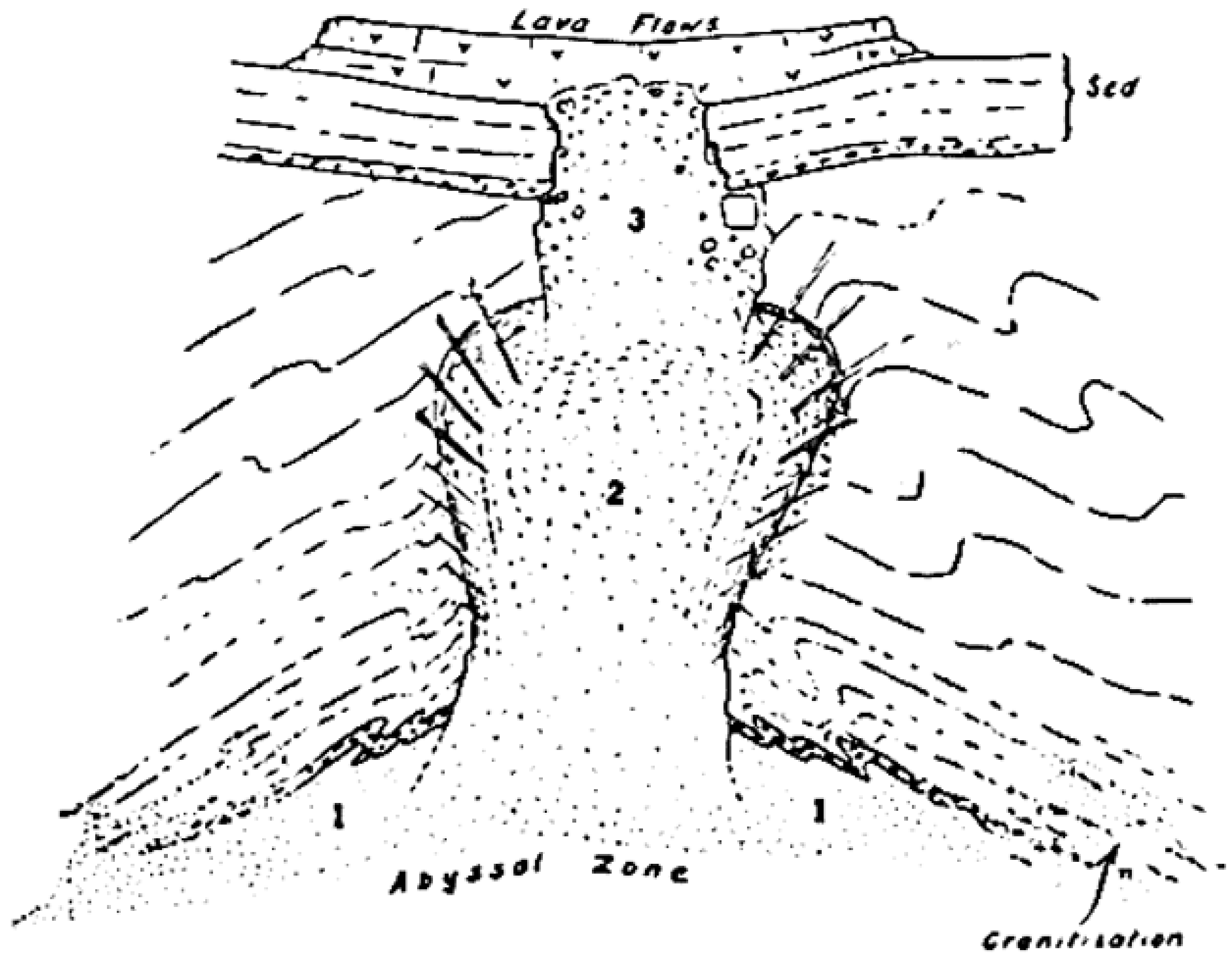


FIG. 98.—DEPTH RELATIONS OF PLUTONIC INTRUSIONS
(Based on H. Cloos, 1931)

1. Abyssal zone, with granitization of the folded sediments, and palingenesis and lit-par-lit injection of the basement on which the sediments rest.
2. Intermediate depth zone, with phenomena of intrusion tectonics.
3. Upper zone, with dominant fracture phenomena, showing magmatic stoping, roof foundering, and early lavas into which the intrusion has subsequently risen.

(2) Large masses of granite, granodiorite, and related rocks are injected into compressed geosynclinal deposits. These intrusions have a maximum development in an intermediate depth zone (the zone of mountain roots), in which pressure and temperature are both high, and most rock types are plastic. It is in connexion with intrusions into fold mountain chains that

[1] See especially papers by Sederholm, J. J., in the *Bull. Comm. Géol. Finlande*, No. 58, 1923: No. 98, 1932: No. 107, 1934. Also van Hise, C. R., 'Principles of North American Pre-Cambrian Geology': *16th Ann. Rept. U.S. Geol. Surv.*, Pt. 1, 1896.

structures included under 'intrusion tectonics' are best developed. The effects of flow together with those of fracture are exhibited by both the intrusion and its wall rocks.

(3) The tectonic features of intrusions in the superficial crustal zones are, especially in the resistant masses, quite distinct from those of the zone of mountain roots and the ultraplutonic zone, owing to the reduction of the plasticity of the country rock consequent upon the lower confining pressure and temperature obtaining there.

Fracture predominates over plastic or viscous flow in the walls of the intrusions, and the border zone structures characterizing the intrusion tectonics of the zone of mountain roots are absent. The roof rocks are dislocated by the rising magma, and the contacts are blocky. Flow structures in the intrusion parallel with its walls are only locally developed, because of the subordination of the effects of diastrophic forces during the emplacement of the intrusions, and the irregular shape of the magma chambers.[1] Roof foundering, accompanied by the extrusion of large volumes of magma along fracture planes, may occur, as in the Erongo district, South Africa.

Structures such as folds and faults may be formed as a direct result of igneous activity, as in the intrusion of laccoliths and plugs, or in the formation of ring dykes and cone sheets. These are all of local significance as regards their details, although the distribution of volcanic and intrusive rocks is, of course, closely bound up with diastrophism. Dyke swarms of regional extent may afford more precise tectonic data, as for example in the marginal flexure of East Greenland and of the Lebombo region of South Africa.[2]

[1] Cloos, H., 'Der Erongo': *Beitr. z. Geol. Erf. d. Deutsch. Schutzgebiete*, Hft. 17, Berlin, 1919; also 'Der Brandberg': *N. Jahrb. f. Min., &c.*, Beil. Bd. 66, Abt. B, 1931, pp. 1–82; 'Granite des Tafellandes und ihre Räumbildung': *ibid.*, Beil. Bd. 42, 1918, pp. 420–55; 'Das Batholithenproblem': *Fortsch. d. Geol. u. Pal.*, Hft. 1, 1923; *Einführung in die Geologie*, Berlin, 1936, pp. 81, 82. Grout, F. F., and R. Balk, 'Internal Structures in the Boulder Batholith': *Bull. Geol. Soc. Amer.*, Vol. 45, 1934, pp. 877–96.

[2] Wager, L. R., and W. A. Deer, 'A Dyke Swarm and Crustal Flexure in East Greenland': *Geol. Mag.*, Vol. 75, 1938, pp. 39–46. Cloos, H., 'Hebung-Spaltung-Vulkanismus': *Geol. Rundsch.*, Vol. 30, 1939, pp. 405–527.

The structures of central volcanic complexes, with which ring dykes and cone sheets are associated in the classic Scottish examples, are of structural interest especially because of the evidence they yield as to the behaviour of the crust under particular stresses. The numerous, thin cone sheets form in families converging downwards, the apices of which lie in circumscribed regions in depth. Each family is believed to arise from the summit of a cupola, from which upward pressure

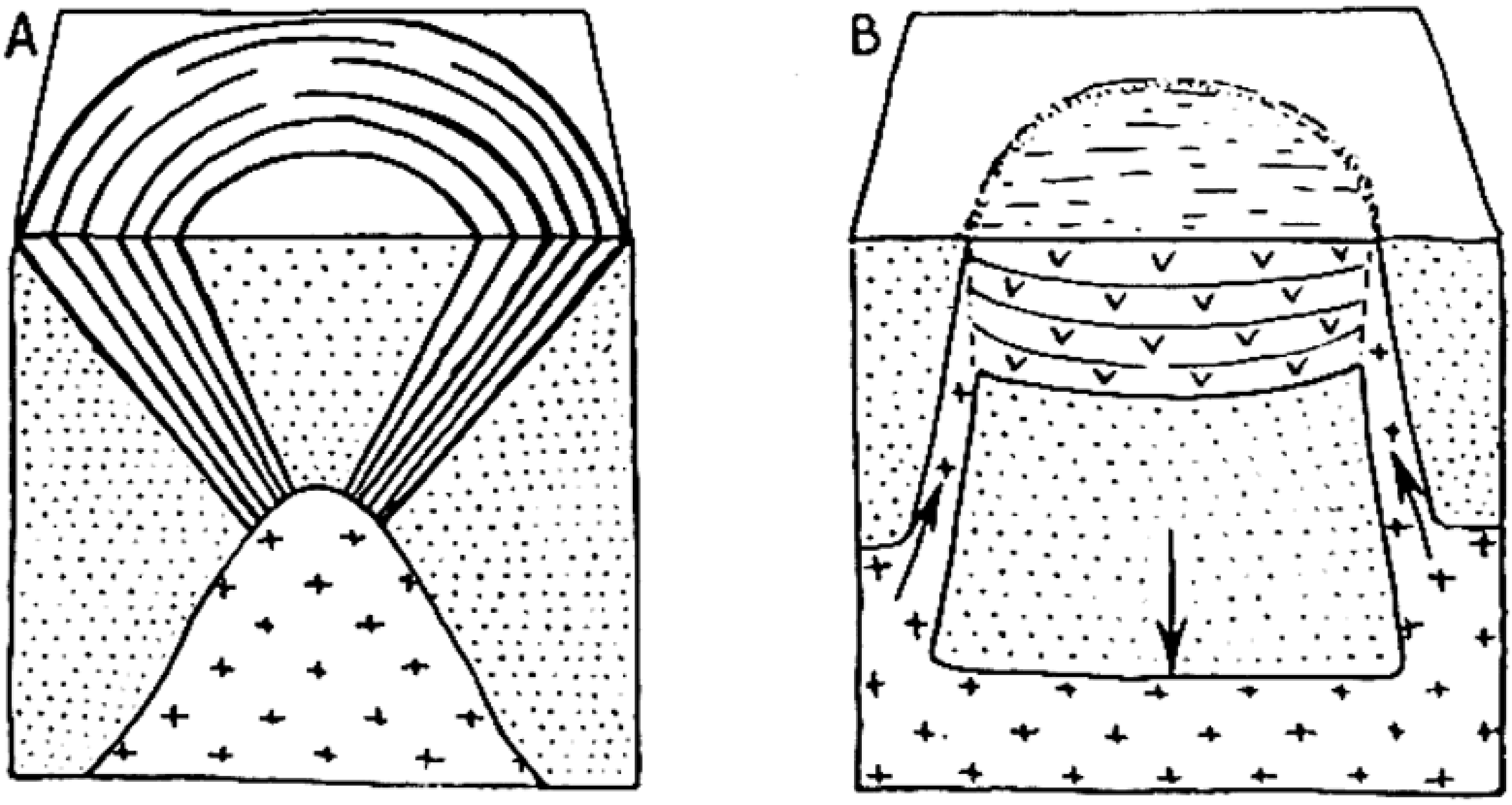


FIG. 99.—CONE SHEETS, RING DYKE, AND CAULDRON SUBSIDENCE

(After Richey)

A. Family of cone sheets arising from the summit of a magma chamber.
B. Cauldron subsidence, with marginal ring dyke and lavas filling the cauldron.

generated the conical fracture surfaces, now occupied by dykes (Fig. 99). Ring dykes, on the other hand, dip vertically or at high angles away from the centre of the complex and represent collapse features formed on relief of pressure from below.[1] Ring and arcuate dykes unaccompanied by cone sheets have been found in many parts of the world, and their recognition, especially of large rings of 10–15 miles across, may involve previously unsuspected correlations of geological data.

[1] Richey, J. E., H. H. Thomas *et al.*, 'The Geology of Ardnamurchan, North-west Mull, and Coll': *Mem. Geol. Surv. Great Britain*, 1930; also 'Scotland: The Tertiary Volcanic Districts': British Regional Geology Series, Geol. Survey, 1935.

Chapter VII

PETROFABRIC ANALYSIS

PLASTIC flow in crystalline materials takes place, as we have noted (p. 33), by displacements of parts of the space lattices of the component grains by twinning and gliding, combined with rotation of the grains relatively to each other. The rearrangements at each part of the deformed body are determined by the stress at that part, and constitute the internal mechanism whereby the body attains the form imposed upon it by the externally applied forces. In addition, the arrangement of new crystals grown during reconstitution under metamorphism is affected by the stress-strain pattern so that by mechanical deformation and the growth of crystals a certain orderliness is induced in the fabric of the rock. Fabric includes crystallographic elements such as optic axes, twin planes and crystal faces, as well as macroscopic structures such as foliation or lineation, and, indeed, refers to any space-data obtainable from a rock. The methods of fabric analysis may therefore be applied to igneous and sedimentary, as well as to metamorphic rocks.[1]

Many significant details may be observed in rocks, apart from full petrofabric analysis of the crystalline components, and their examination is often practicable in the field. These include deformed fossils, pebbles and oolite grains, and small-scale structures such as lineation, puckered and sheared

[1] The following are standard works: Sander, B., *Einführung in die Gefügekunde der Geologischen Körper*, Bd. I, 1948, Bd. II, 1950, Vienna. Schmidt, W., *Tektonik und Verformungslihre*: Berlin, 1932. Knopf, E. B., and E. Ingerson, 'Structural Petrology': *Mem. Geol. Soc. Amer.*, No. 6, 1938. Fairbairn, H. W., *Structural Petrology of Deformed Rocks*: Cambridge, Mass., 1949. See also Demay, A., 'Microtectonique et Tectonique Profonde': *Mem. Carte Géol. France*, 1942.

laminae, augen, and so on,[1] examples of which are shown in Fig. 107. Use has also been made of 'pressure shadows'—augen-like growths of secondary quartz and other minerals, that often occur around hard crystals such as pyrites or magnetite in slates and schists, and whose long axes are taken to correspond with the direction of elongation in the rock.[2]

Lineation, which has been defined as 'any kind of linear structure within or on a rock', has been widely used in tectonic analysis, but since the term includes structures of widely diverse origin its use must be guided by previous interpretation of its structural significance in the area under investigation. The importance of lineation lies in the evidence it affords of the direction of movement. In one class, for example in slickensides, lineation is parallel with the movement in a given plane; in another, as for instance with the lines representing the intersection of cleavage and bedding, it is normal to the direction of movement and parallel with the tectonic axes of the folds. The subject has been fully discussed by E. Cloos.[3]

Tectonites and Non-Tectonites.—Intragranular lattice displacement, grain rotation, and ionic migration by diffusion or solution, which take place without destroying the cohesion between the grains of a deformed body, are termed *componental movements* (*Teilbewegungen*), and *tectonites* are defined as rocks which have undergone componental movements. The fabric of tectonites is referred to as *deformational fabric*. All other rocks

[1] An excellent summary is given by E. Cloos, 'Oolite Deformation in the South Mountain Fold, Maryland': *Bull. Geol. Soc. Amer.*, Vol. 58, 1947, pp. 843–918.

[2] This interpretation is implicit in the German and French equivalents—*Streckungshöfe*, *halos d'etirement*, and is adhered to herein although Mügge has argued that they are formed by shearing movements. Such movements play a part, although subordinate, in some examples (Fig. 107). See Mügge, O., 'Bewegungen von Porphyroblasten in Phylliten und ihre Messung': *Neues Jahrb. f. Min.*, *&c.*, Vol. 61, Beil. Bd. A, 1930, pp. 469–510. Bain, G. W., 'Wall-Rock Mineralization along Ontario Gold Deposits': *Econ. Geol.*, Vol. 28, 1933, pp. 705–43. Pabst, A., '"Pressure-Shadows" and the Measurement of the Orientation of Minerals in Rocks': *Amer. Min.*, Vol. 16, 1931, pp. 55–70.

[3] Cloos, E., 'Lineation': *Geol. Soc. Amer.*, Mem. 18, 1946.

are classed as *non-tectonites*, but some of these, as for example sediments containing mica flakes lying parallel to the stratification planes, may possess an orientated micro-fabric simulating that of tectonites. Orientated fabrics resulting from the settling of grains through fluid media are termed *depositional* (*Anlagerungsgefüge*). *Growth fabric* develops by crystallization of minerals in place, as in the parallel growth of quartz in many veins, but in tectonites there is a complex interrelationship between growth and deformational fabrics, which is not fully understood.

Trener,[1] who was the first to carry out the petrofabric analysis of a tectonite, found that a majority of the quartz crystals in a dynamically metamorphosed quartzite from Tonale were arranged with their optic axes perpendicular to the schistosity, exhibiting a *preferred orientation*. Fabric analyses may now be accurately made with the universal microscope stage, but for fine-grained rocks X-ray methods must be used, and for the most part investigations have been restricted to medium-grained rocks.

Development of Preferred Orientations.—In homogeneous non-crystalline or extremely fine-grained materials, plastic flow occurs by means of internal shearing movements along certain planes whose attitude is determined by the stress and the physical properties of the material (pp. 33–8). When plastic flow commences, those planes along which the slip involves the least work will become shearing planes. In crystals, the regular atomic packing causes the physical characters to be vectorial, and shearing movements occur preferentially along certain planes called *glide planes*. Slip along these planes is of such a nature that the displaced portions of the crystals, after gliding, fit together and re-establish the lattice structure across the glide planes, cohesion thus being maintained. Not only does gliding take place along certain preferred crystallographic planes, however, but it is also restricted, on the whole, to a particular direction, called the *glide direction*, in those planes. The net or statistical result of the existence of preferred glide directions in the individual mineral grains in a rock is the development in

[1] Trener, G. B., 'Geologische Aufnahme an nördlichen Abhange der Presanellagruppe': *Jahrb. d. Geol. Reichsanst.*, 1906, pp. 453–70.

fabric as a whole of a *glide line*. Displacements of the above type in crystal lattices are referred to as *translation gliding*. Twinning, too, is restricted to lattice rotations about certain axes, which results in a gliding movement (*twin gliding*) parallel to the twin plane. Thus, during plastic flow of crystalline materials, the shearing planes do not cut indiscriminately across the grains, but each crystal grain yields along predetermined directions of ready displacement that are characteristic for each crystalline substance.[1]

If a crystal grain is so orientated that the position of its gliding or twinning planes is sufficiently close to the direction of the planes of potential shear which are generated by the stress, it will yield along those gliding or twinning planes. If not, it will rotate as a whole until such intragranular movements can occur, or until it is prevented by the surrounding grains from rotating any further. Again, glide planes developed in a 45° position to the principal stress axes rotate during deformation and assume an attitude more nearly parallel with the axis of elongation. Deformed crystalline substances therefore exhibit *preferred orientation* of crystallographic fabric elements, determined by the stress and the crystallographic properties (e.g. direction of twin and glide planes and ease of lattice displacement in these) of the grains.

Preferred orientation of lattice elements is termed *lattice orientation*, and that which is revealed by the external form of the grains, *dimensional orientation*. The latter results in part from the tendency of elongated or platy crystals to rotate into parallelism with the shearing planes in solids and the flow lines in liquids, and in part from other causes (see, for example, pp. 160–1). Grains showing no dimensional orientation, as in granular aggregates of quartz in a schist, may nevertheless exhibit strong preferred lattice orientation. On the other hand, parallel quartz fibres show marked dimensional orientation, although their crystallographic directions may have no preferred orientation.

[1] A summary of theories concerning the mechanism of gliding and twinning is given by Burgers, in Houwink, R., *Elasticity, Plasticity and Structure of Matter*: Cambridge, 1937, Chap. 5.

Analysis of Motion in Tectonites.—In order to enable correlation to be made between microscopical fabric orientation and macroscopic structures, all directional structures in the rock are described with reference to three rectangular co-ordinates labelled the *a*, *b*, and *c* axes, as in crystallography: *a* is the direction of motion in the rock, and may be either a direction of shearing movement or of compression in folding; *b* is

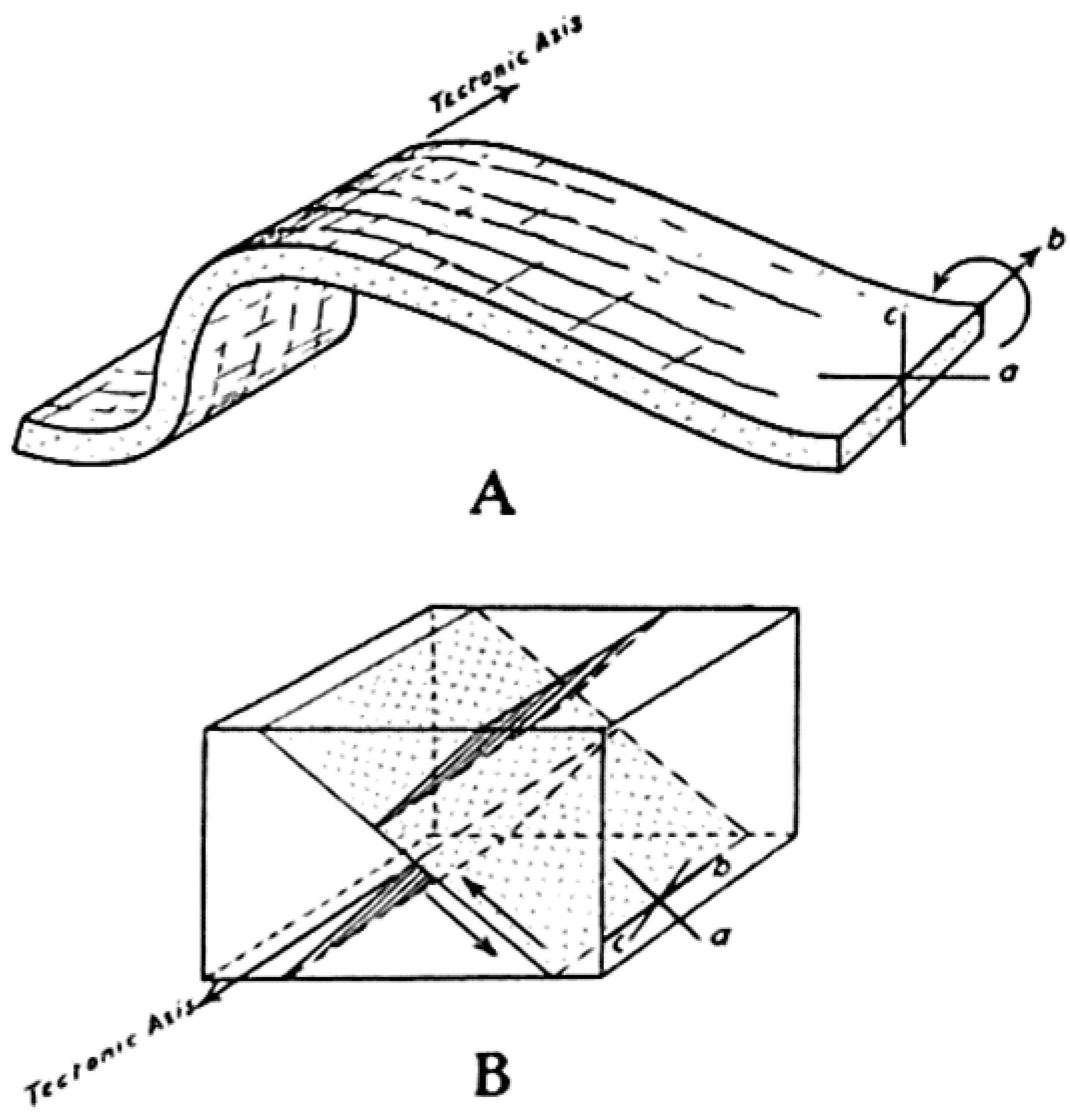


FIG. 100

A. Orientation of the fabric reference axes *a*, *b*, and *c*, in folding. Bedding plane slip causes grain rotation about the tectonic axis (*b*).

B. Orientation of fabric reference axes for the stippled shearing plane, which is of later origin than the displaced shaded plane.

normal to the direction of motion, lying in the plane of slip (bedding plane slip or slip along a shearing plane). The *b* co-ordinate is a direction that is often visible macroscopically in the rock, as a lineation on planes of foliation, minute puckering, or as the intersection of two complementary shearing surfaces developed by the same deformation. In buckle folds it corresponds with the crests and troughs, and it may be termed the *tectonic axis*. The *c* co-ordinate is the normal to the *ab* plane. For an example illustrating the choice of axes, see pages 162–3,

and Fig. 100. As in crystallography, the intercepts of any plane on the *a*, *b*, and *c* axes are denoted by *h*, *k*, and *l* respectively. An *hol* plane, for example, is parallel to the *b* axis and cuts the other two. If the three axes of a triaxial strain ellipsoid are denoted as A, B, and C, of which A is the long, C the short, and B the mean axis, then the *b* co-ordinate of the fabric is parallel to the B axis of the strain ellipsoid. If the rock has been deformed by forces acting from different directions at different periods, each deformation is referred to a different set of co-ordinates called *a′*, *b′*, *c′*, and so on.

In practice, either one or two linear elements (lineations) are usually visible in a rock, and in some instances as with slickensides, it is clear that the main lineation is in *a*.

In many folded rocks it is assumed that the direction of transport *a* is normal to the axial planes, and that *b* is the tectonic axis of the folds.

This would not apply in shear folding, however, and, in fact, the physical significance of tectonite planes and axes is not always clear.[1]

S-surfaces.—In fabric analysis, the recognition of planar elements is fundamental, and Sander has called all such planes *S*-surfaces. These may be visible as individually distinct planes such as bedding planes, cleavage planes, faults, or fractures ; they may be visible but not individualized, as in a schist, when the planar schistosity permeates the whole rock; or they may be without visual expression, as in the 'statistical' planes afforded by the parallel orientation of crystallographic elements revealed by the preparation of petrofabric diagrams (see pp. 162–3). Tectonically developed *S*-surfaces are controlled by the stress-strain pattern to which the rock has been subjected, but the relationship is not simple and its interpretation is generally controversial, as is shown by the long-continued disputation on the relationship of axial-plane cleavage or schistosity to the strain axes. According to Becker and

[1] For critical discussion of such problems see Anderson, E. M. 'On Lineation and Petrofabric Structure etc.': *Quart. Journ. Geol. Soc.*, Vol. 104, 1948, pp. 99–132. Knopf, E. B., 'The Record of Deformation Movements shown by Petrofabric Analyses': *Amer. Journ. Sci.*, Vol. 241, 1943, pp. 336–42.

others[1] flow cleavage represents shear planes formed obliquely to the direction of compression, whereas van Hise, Leith, Heim[2] and others maintain that it forms in the AB plane of the strain ellipsoid, normal to the compression.

It is probable that there is no general solution of this problem, since several factors are involved and these will vary in different environments. Firstly, lamellar particles embedded in compressible material are rotated during deformation, and with strong compression or elongation they come to lie statistically parallel with the AB plane of the deformation. This *external rotation* of the grains results in dimensional orientation; similar rotation of grains growing during deformation results in both dimensional and lattice orientation, but the orientation of such grains may be related to shearing planes rather than to the plane of flattening.

Dimensional grain orientation may develop in the AB plane even in rocks of relatively low ductility, as shown by Sander.[3] The equal development of complementary sets of shearing planes operating in pebbles in conglomerates or in mineral grains results in elongation of these fabric elements in a plane normal to the pressure (see Fig. 101). The visible effect is a schistosity parallel to this plane, which may be termed the *plane of flattewing* (*Plättungs-s*)[4] since the dimensional orientation of the grains in the plane of flattening arises by means of lattice deformation, it is termed pseudo-dimensional orientation. Fairbairn[5] has emphasized that elongation of grains parallel to the A axis of the strain ellipsoid will take place only if the grains do not rotate through any appreciable angle during the deforma-

[1] Becker, G. F., 'Schistosity and Slaty Cleavage': *Journ. Geol.*, Vol. 4, 1896, pp. 429–48; 'Current Theories of Slaty Cleavage': *Amer. Journ. Sci.*, Vol. 24, 1907, pp. 1–17.

[2] van Hise, C. R., 'Principles of North American Pre-Cambrian Geology': *16th Ann. Rept. U.S. Geol. Surv.*, Pt. 1, 1896, p. 639. Leith, C. K., 'Rock Cleavage': *U.S. Geol. Surv.*, Bull. No. 239, 1905. Heim, A., *Geologie der Schweiz*, Bd. II, 1925.

[3] *Gefügekunde*: pp. 219–20.

[4] The term *plaiting surface* has been used by Fairbairn and some later writers as an equivalent for *Plättungs-s*, but is now regarded as unsatisfactory.

[5] Fairbairn, H. W., 'Elongation in Deformed Rocks': *Journ. Geol.*, Vol. 44, 1936, pp. 670–80.

tion. If rotation occurs, a particular dimensional axis in a pebble or mineral grain will at one moment be parallel to A, and therefore elongated, while later it will rotate away from A, and thus be less elongated or even shortened. The B axis of a triaxial strain ellipsoid is, moreover, a direction of subsidiary elongation, and if the effects of elongation parallel to A are inhibited by grain rotation, B becomes the dominant direction of elongation of fabric elements, and the rock develops a linear fabric parallel to this axis. This effect, however, will usually be noticeable only when considerable grain rotation occurs (see pp. 26–8). Elongation of fabric elements along the B axis of the strain ellipsoid may cause the formation of tension gashes in the AC plane, these becoming filled with newly formed minerals, and imparting a well-defined schistosity to the rock.

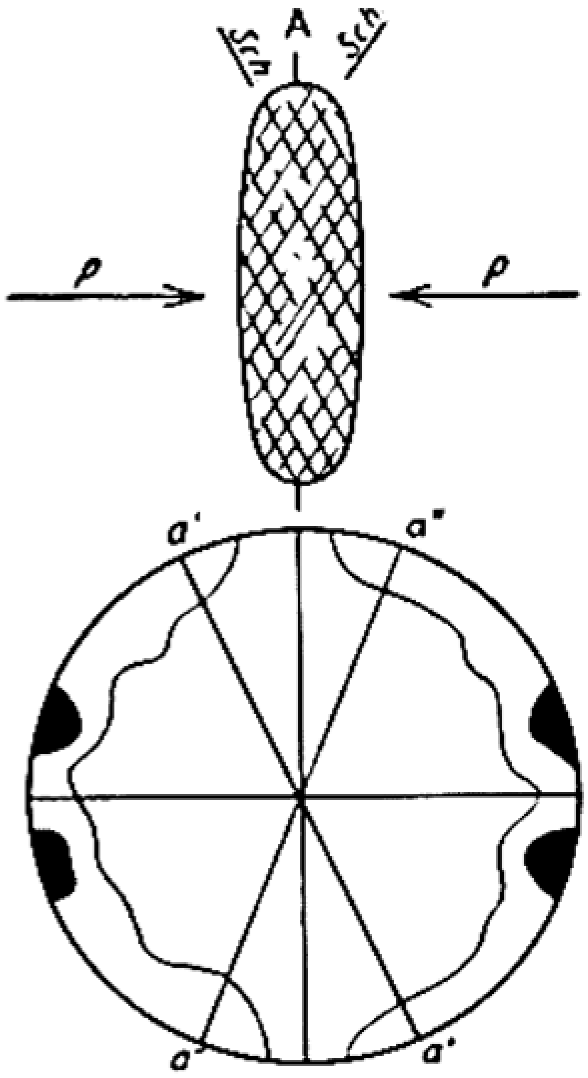


FIG. 101.—A PARTICLE SUBJECTED TO PRESSURE P DEVELOPS SHEARING PLANES SCH., WHICH ALLOW IT TO ELONGATE ALONG THE AXIS A

(After Fairbairn and Sander)

A sketch orientation diagram in the AC plane of the strain ellipsoid (the plane of the paper) shows the shearing planes *a*, *a'*, with maxima related to them.

Again, orientated fabric may result from the crystallization of new minerals along pre-existing *S*-surfaces, including bedding (as in bedding schistosity), shear planes, tension cracks and so on. This is *mimetic crystallization*, and if it takes place after deformation is complete it yields dimensional orientation only; but if such growth is affected ever by small residual stresses, lattice orientation may well result. Finally, experiments show that in compressed heated glass, lamellar crystals grow with their flat faces normal to the pressure.[1]

[1] Wright, F. E., 'Schistosity by Crystallization—A Qualitative Proof': *Amer. Journ. Sci.*, Vol. 22, 1906, pp. 224–30.

The difficulty of distinguishing the effects of rotational and irrotational strains must also be considered (see pp. 26–7), and finally, the physical state of the rock itself. Although generally considered as solids, slates and schists show evidence of the presence of a fluid phase, some of which is produced in slates by the release of water from clay minerals when they re-crystallize to mica. Several authors have commented on the apparent necessity of considering the possible existence of planes comparable with fluid flow planes in such rocks.

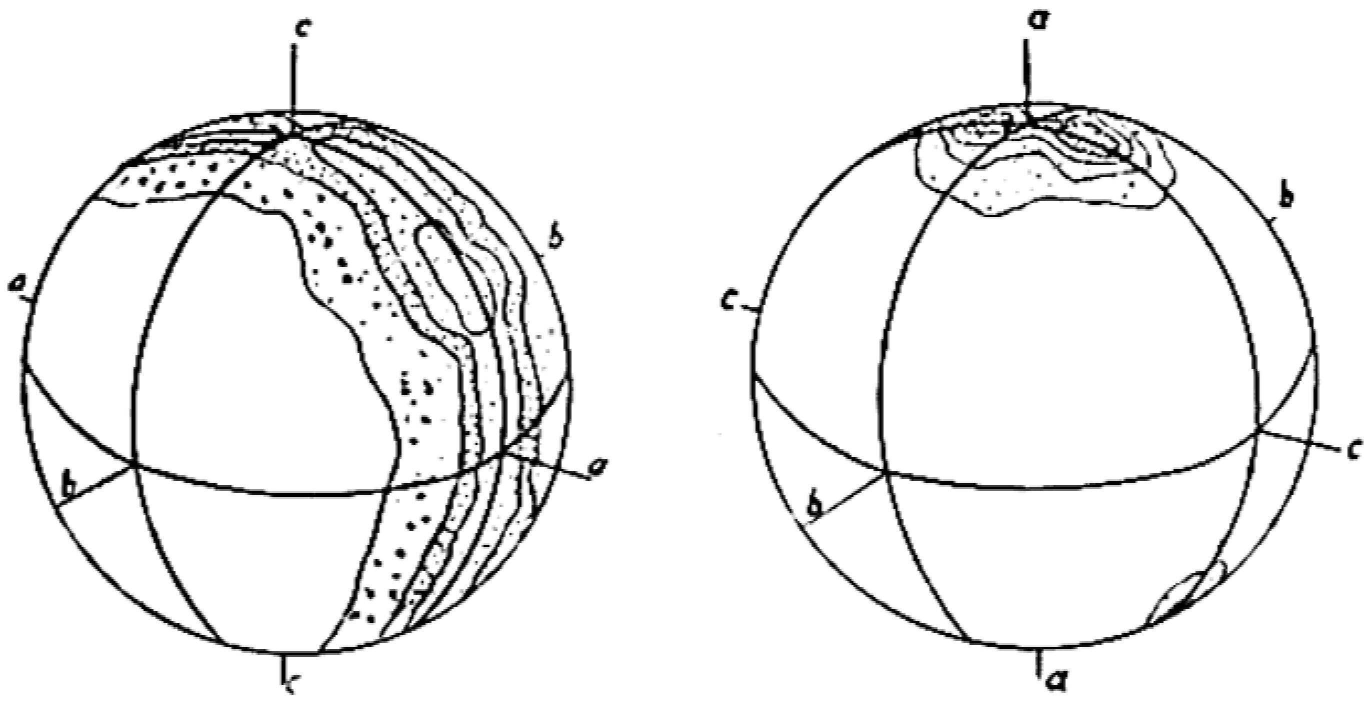
Method of Procedure in Petrofabric Analyses.—In order that the orientation of macroscopic and microscopic fabric elements may be related to the tectonics of the deformed rocks as a whole, it is necessary to collect hand specimens with care, choosing examples in which the various structures are well shown, and orientating them in the field by marking the specimens so that they may be re-orientated in the laboratory parallel to their original position in space.

In the laboratory study, the most prominent plane structure of tectonic origin, such as a well-developed foliation or cleavage, is provisionally taken as the *ab* plane. The intersection of complementary slip surfaces being the B strain axis, which corresponds with the *b* space reference axis, this direction is then determined by noting lineation on the *ab* plane. Such lineation is generally caused by the intersection of a set of slip planes or a plane of flattening with the *ab* plane. However, since lineation may in certain circumstances form in *a*, and since girdles may also form about the *a* axis, the determination of the true B-axis is often a matter of some difficulty, requiring the correlation of regional observations with laboratory data.

The direction in which thin sections will be cut is determined by the choice of the reference axis, three sections generally being cut, normal to the *a*, *b*, and *c* axes respectively. If more than one set of axes is recognized, sections are also cut normal to them. Even in rocks such as some fine-grained marbles, in which the macroscopic tectonic structures are not present, fabric orientations may be found, and the axes are then determined by first cutting random sections in order to discover the

orientation. Before proceeding to the microscopic examination, the thin sections are examined with a binocular microscope, and all structures visible are represented in sketches. Under the binoculars, the direction of slip along the various shearing planes, and also the relative ages of the sets of slip planes, can sometimes be determined by noting the orientation of platy mineral flakes. Finally, the orientation diagrams are prepared.

The statistical investigation of the orientation of crystallographic fabric elements is carried out with the universal stage attached to a petrological microscope. The poles of optic axes,



A B

FIG. 102.—SPHERICAL PROJECTIONS SHOWING THE POLES OF FABRIC ELEMENTS IN A B-TECTONITE (A) AND AN S-TECTONITE (B)

The projections are contoured to illustrate the meaning of the contours on the ordinary orientation diagrams, which are equal area projections of the lower halves of the spherical projections.

twinning planes, and crystal faces are located on a spherical projection (see Fig. 102) and the results recorded on the Schmidt net, which is an equal-area azimuthal projection of the lower hemisphere of the spherical projection. Areas in which the density of spacing of the poles falls within chosen limits, e.g. 1, 2, 3, 4 per cent, &c. of all grains examined falling within 1 per cent of the area of the projection, are then determined, and contours drawn as boundaries to them. The area of greatest crowding is indicated clearly, generally by solid black, and the reference planes *ac*, *ab*, and *bc* are marked on the projection (see Fig. 105).

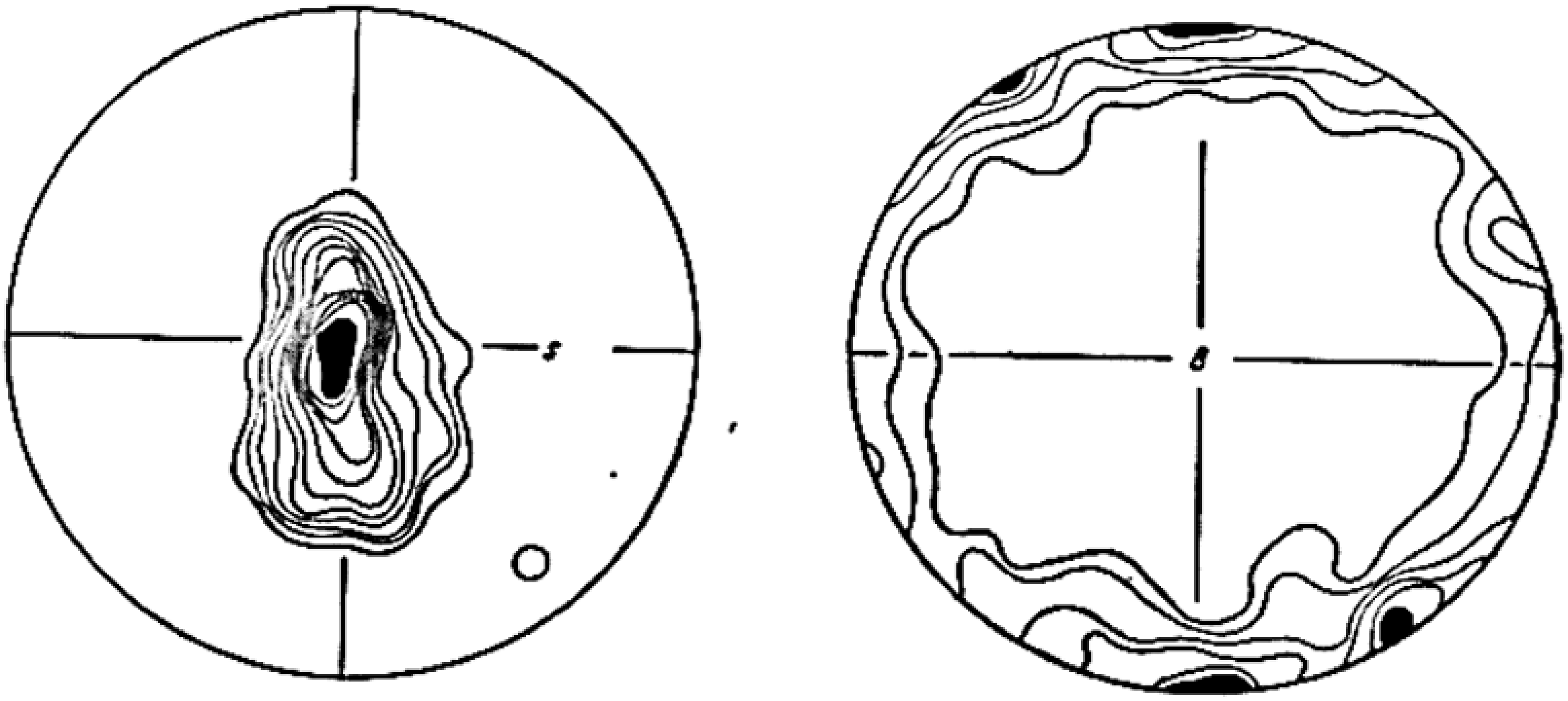
Types of Tectonites.—Two contrasted types of tectonites are revealed by orientation diagrams and macroscopic study of rocks. In the first there are one or more well-defined maxima, indicating that the development of the fabric orientation was controlled by *s*-surfaces, but there is no linear element parallel to the *b*-axis. Such rocks are therefore termed *S-tectonites* (see Figs. 102 A, 103 (1)). In the second type, the poles of the crystallographic elements fall within a girdle round the spherical projection, with or without maxima, indicating that the dominant control of fabric orientation was effected by rotation about an axis normal to the girdle, or by shearing planes intersecting along that axis. The most important axis of rotation during deformation being the *b*-axis, such rocks are termed *B-tectonites*, and the *b*-axis in them is labelled B. If rotation about B has been dominant, the rocks are termed *R-tectonites*, and every gradation exists between S-, B-, and R-tectonites (see Figs. 102, B, 104 (2)). The characteristic girdle of R-tectonites develops when the deformation is prolonged after a majority of the grains have assumed an orientation favourable to gliding or twinning. When numerous lattice displacements have taken place, further gliding or twinning becomes difficult, as in the work hardening of metals, and external rotation may become easier than lattice displacement. The grains then rotate in the *ac* plane and produce the girdle.

Experience has shown that the most satisfactory minerals from which to construct orientation diagrams are quartz, calcite, and the micas. In quartz, the optic axes are mapped; in calcite, optic axis, and the normal to the twin and glide plane ($01\bar{1}2$); in mica, generally the normal to the (001) plane.

Symmetry of Orientation Diagrams.—A glance at a number of orientation diagrams (see Figs. 102, 103) will show that they present a certain symmetry. This is even better revealed by forming a mental picture of the spherical projection from which the diagrams are derived. The following types of symmetry are recognized.[1]

[1] Sander, B., *Gefügekunde*: p. 146.

1. Isotropic: completely haphazard arrangement of fabric elements (Fig. 104 (1)).
2. Spheroidal: the spherical projection has the symmetry of a spheroid of revolution (Fig. 103 (1)).
3. Orthorhombic: the spherical projection has the symmetry of a triaxial ellipsoid. There are two planes of symmetry in the orientation diagram (Fig. 104 (2)) the other being the plane of the diagram.



1 2

FIG. 103.—ORIENTATION DIAGRAMS OF TYPICAL S- AND B-TECTONITES

(After Sander, *Gefügekunde der Gesteine*)

1. S-tectonite; optic axes of quartz grains in a mylonite band in the Melibokus granite. S is the plane of shearing.
2. B-tectonite with subsidiary maxima and minima; poles of the (0112) twin and glide plane in calcite. Calcphyllite, Brenner, Tirol.

4. Monoclinic: the spherical projection has a centre of symmetry, one plane of symmetry, and one axis of symmetry at right angles to the plane. There is one plane of symmetry in the orientation diagram (Fig. 104 (3)).
5. Triclinic: the spherical projection has only a centre of symmetry. The orientation diagram has no symmetry (Fig. 104 (4)).

The symmetry of the fabric orientation is related to the partial movements in the rock. Spheroidal symmetry may

express the absolute dominance of one *s*-surface, with no visible *b*-axis. The optic axes of the grains in the quartz films on slickensided surfaces show this orientation, and these quartz films also afford a good example of an S-tectonite (see Fig. 103 (1)). The fabric orientation diagrams of lamellar grains

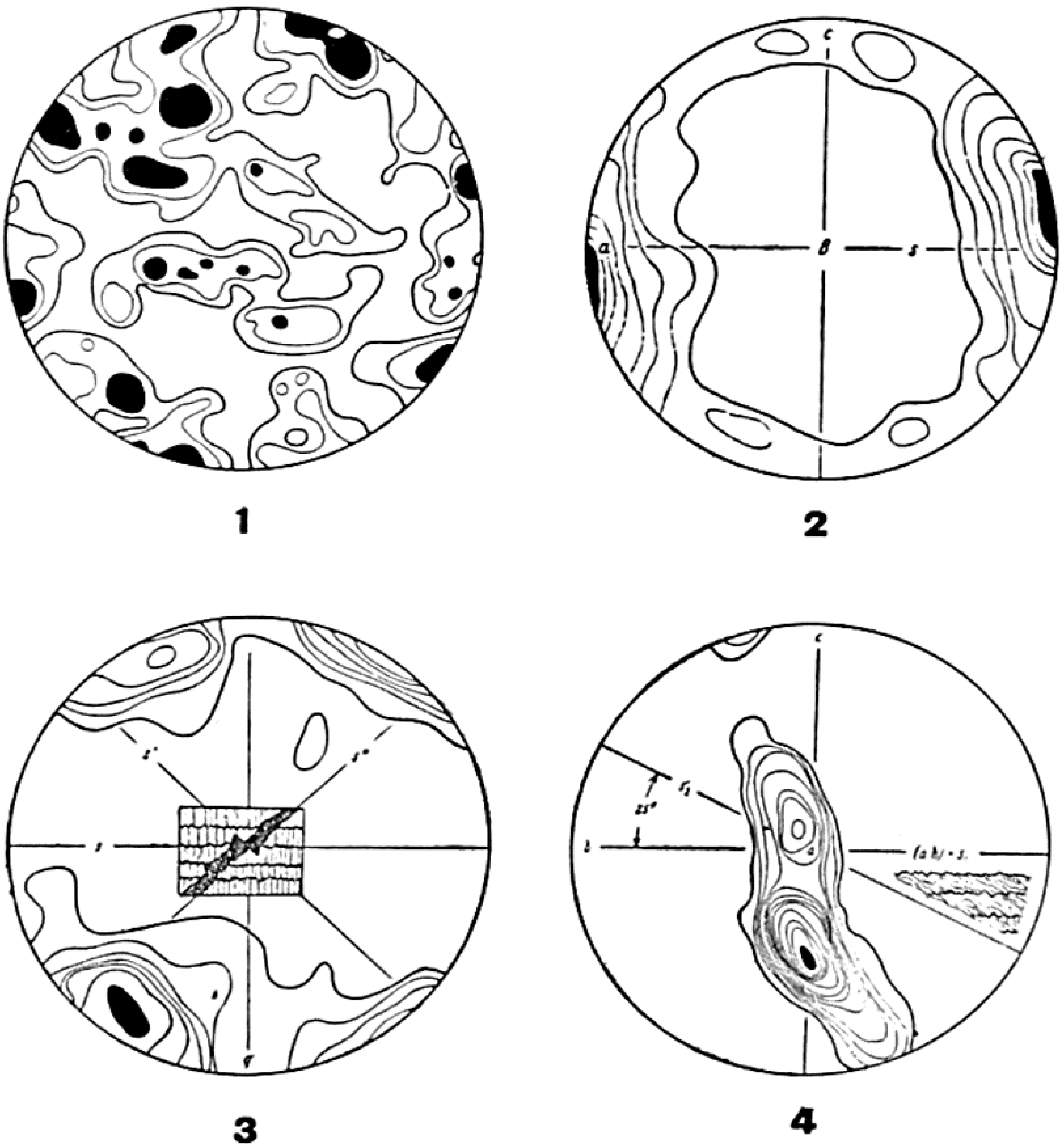


FIG. 104.—ORIENTATION DIAGRAMS SHOWING VARIOUS SYMMETRY TYPES

(After Sander, *Gefügekunde der Gesteine*)

1. Isotropic; poles of (010) faces of andesine in orbicular diorite.
2. Orthorhombic; poles of (001) faces of biotite in biotite schist.
3. Monoclinic; optic axes of quartz crystals in the shear plane S″, traversing the strongly orientated crystals whose long axes are parallel to *q*.
4. Triclinic; optic axes of quartz in thinly foliated pegmatite.

that have settled through a fluid medium also show spheroidal symmetry. Orthorhombic symmetry in tectonites expresses the presence of two equally well-developed *s*-surfaces, generally intersecting in *b*.

Monoclinic symmetry is commonest in tectonites. It is the symmetry of a forward rolling motion like that of a wheel, and is perhaps produced in tectonites by the rotational movements of the grains. Typical R-tectonites show monoclinic symmetry, the symmetry plane being the *ac* plane, and the axis the *b*-axis.

Triclinic symmetry is shown by tectonites having intersecting *hol okl* and *hkl s*-surfaces, due either to deformation at different times by forces acting along different directions, or to rotation about an axis other than the *b*-axis.

Although the recognition of symmetry-types is useful, it is found in practice that numerous deviations from the idealized types occur, which necessitate a broad rather than a precise view to be taken. The orientation diagram itself has to be interpreted in the light of the lattice translations that are known or believed to take place in the minerals examined, the collective data regarding the movements in the rock, and the presumed effects these movements would have on lattice orientation.

Recrystallization in Relation to Fabric.—The microfabric which will result from deformation of a certain kind will obviously be affected by recrystallization of the component minerals of the rock. Although at first sight it might be thought that crystallization of post-deformation date (*post-tectonic crystallization, pre-crystalline deformation*) would obliterate all the directional structures resulting from the strain, this is not the case. *S*-surfaces are planes which afford a ready passage for solutions, and along which the resistance to crystal growth is less than in the body of the rock. Thus, on recrystallization, pre-existing *s*-surfaces are outlined with newly formed minerals generally of lamellar habit, and arranged with their flat surfaces parallel to the *s*-surfaces. This process is called *mimetic crystallization* (*Abbildungskristallisation*), and is one method whereby dimensional grain orientation giving a schistose structure to the rock may be developed.

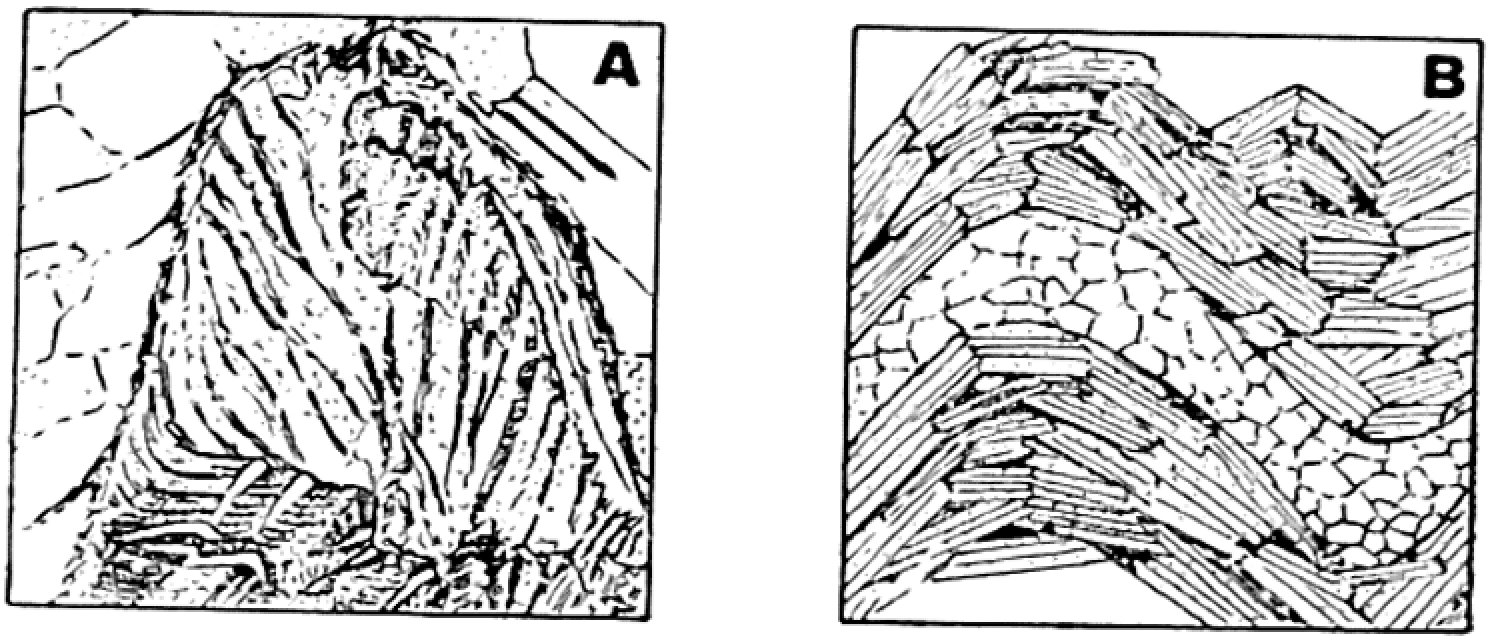


FIG. 105.—A. POST-CRYSTALLINE DEFORMATION, SHOWN BY DISTORTED MICA CRYSTALS IN THE CORE OF A FOLD IN GNEISS. B. POST-DEFORMATION CRYSTALLIZATION SHOWN BY UNDISTORTED MICA FLAKES IN MINUTE FOLDS IN MICA SCHIST. THE QUARTZ LAYER SHOWS POST-CRYSTALLINE DEFORMATION

In *A* the crystals are not distorted, and grew after the folding; in *B* they are all distorted by the folding.

With *para-crystalline deformation* (*syntectonic crystallization*) deformation and recrystallization go on at the same time, and if the deformation follows recrystallization, it is termed *post-crystalline deformation* (*pre-tectonic crystallization*).

The time relationships of recrystallization and deformation are best revealed by minerals which react readily to stress, such as mica, quartz, calcite, and gypsum. The mica flakes on

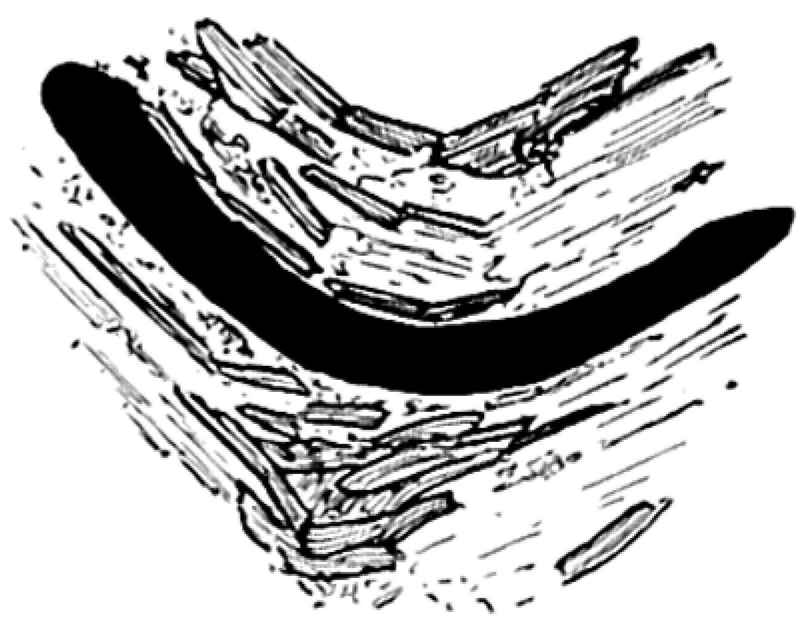


FIG. 106.—PARA-CRYSTALLINE DEFORMATION SHOWN BY THE DISTORTION OF THE MICA FLAKES BELOW THE BENT PLATE OF IRON ORE, AND THE LACK OF DISTORTION IN THE CORE OF THE FLEXURE (× 15 APPROX.)

(Diagrammatic, after Sander)

both the inner and outer sides of minute folds in a puckered schist, sometimes show well-developed crystal outlines and lack of distortion (Fig. 105, B). This indicates pre-crystalline deformation and mimetic crystallization. In other rocks, the cores of the minute folds contain undeformed mica flakes,

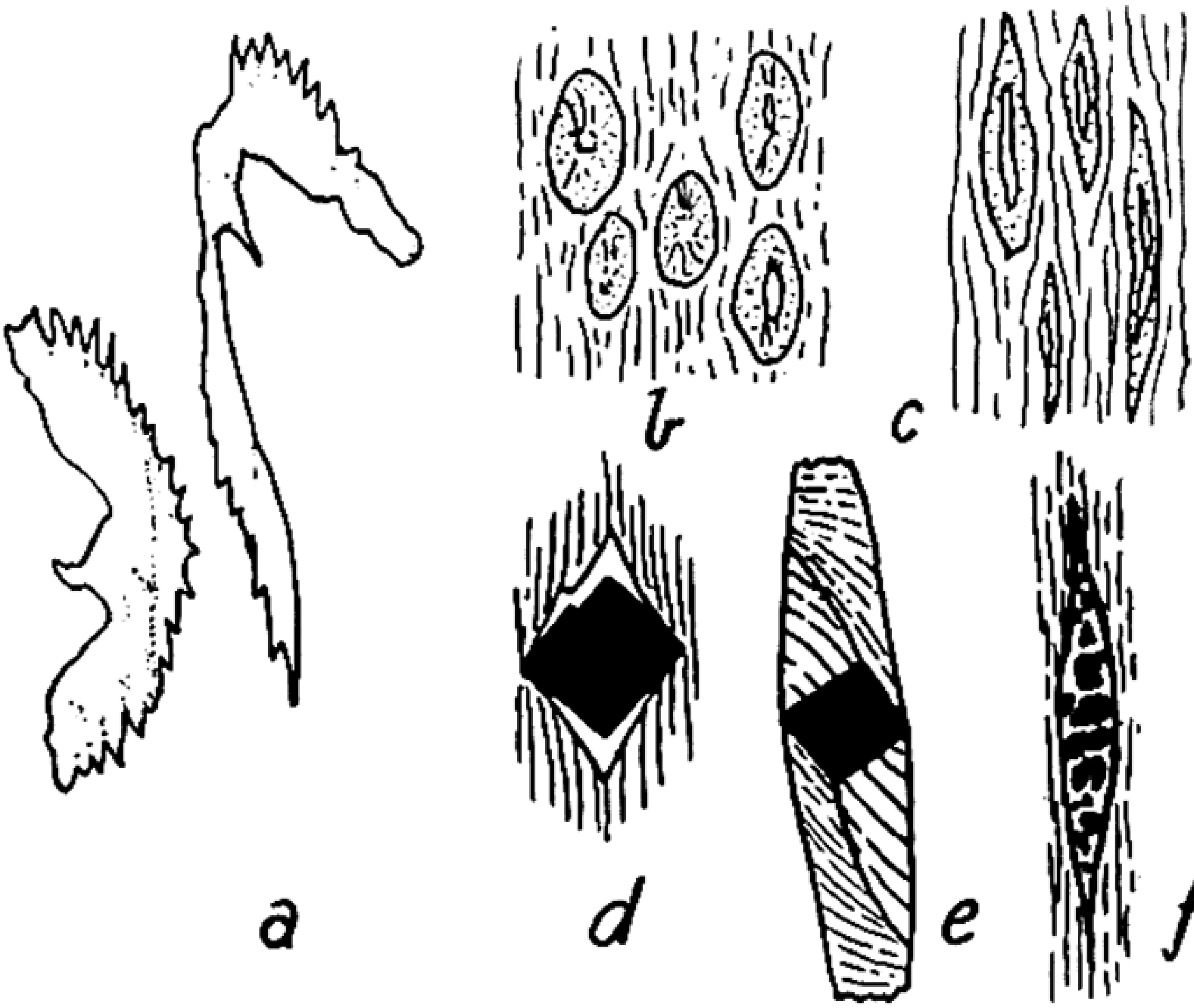


FIG. 107.—DEFORMATION OF OBJECTS IN ROCKS

(Cleavage is vertical in all instances.)

(*a*) Graptolites (*Isograptus caduceus*), showing compression normal to cleavage.

(*b*), (*c*) Deformed ooids in oolitic limestone. A deformation 50%; Po *b*, 100% (after E. Cloos).

(*d*), (*e*), (*f*). Pressure fringes around pyrites in slates.

a—elongation in the cleavage direction, no rotation;
b—the same, combined with rotation (after Mügge);
c—disrupted pyrites grain in pressure-fringe (after Magnée).

although the flakes on the outer sides of the flexures are distorted (Fig. 106); or again, the mica flakes in one fold will be deformed, and in another well crystallized, with no evidence of distortion. These relationships indicate syntectonic crystallization of the mica. This is also indicated for the calcite porphyroblasts in the slate shown in Plate III, Fig. B. It will be noted

that the lines of inclusions in the calcite, which represent residual grains originally parallel to the present cleavage direction, are sigmoidally curved. As the porphyroblasts grew, they were at the same time rotated by shearing movements along the cleavage planes, so that the lines of inclusions in successive layers from the inside of each crystal towards its periphery, show progressively less rotation.

Pre-tectonic crystallization of a particular mineral is indicated by cataclasis of all the grains (see Fig. 105, A). The different minerals in a metamorphic rock do not all develop at the same time, however, and thus the time relationships between crystallization and deformation will not be the same for each species.

Examples of the application of the methods of petrofabric analysis will be found in the works listed below.[1]

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